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VOLUME LI NUMBER 3

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APRIL-MAY 1943

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THE JOURNAL OF GEOLOGY

April-May 1943

CRITICAL STRUCTURES OF THE GROS VENTRE AND NORTHERN HOBACK RANGES, WYOMING

VINCENT E. NELSON
University of Kentucky
AND
VICTOR CHURCH
Utah State Agricultural College

ABSTRACT

The Rocky Mountain geosyncline had an irregular development, so that locally there were basin-like areas where pre-Laramide sediments accumulated in greater thicknesses than elsewhere. This marked difference in total thicknesses of Paleozoic and Mesozoic sediments gave rise to two distinct types of structural development in the Middle Rocky Mountain province, namely, foreland and geosynclinal. The area discussed in this report is particularly significant from the standpoint of the mechanics and sequence of development of these structural types, for only at the southern end of the Teton Range and at the junction of the Gros Ventre and Hoback ranges are pronounced structures of the two types in juxtaposition. The structures of the western and central Gros Ventres are described and shown to be typical of a foreland area, while the northern end of the Hoback Range displays structures typical of a geosynclinal tract. Laramide deformation in this border zone occurred in two stages, with the Gros Ventres (foreland area) yielding first and serving as a buttress against which the structures of the geosynclinal tract were crowded. A period of Tertiary high-angle faulting is also recorded, together with comparatively recent epeirogenic movements.

INTRODUCTION

This report is based on field work carried out by the writers during portions of the summers of 1937, 1938, and 1939. Mapping for the study covers an area of approximately 240 square miles, lying within the Jackson and Gros Ventre quadrangles of northwestern Wyoming. It includes the central and western portions of the Gros Ventre Range and the northern end of the Hoback Mountains.

The Gros Ventre Mountains constitute one of the more or less detached ranges of the west-central portion of the Middle Rocky Mountain province. They are a relatively small unit, being but 40 miles long and from 15 to 20 miles wide and trending northwest southeast. The western termination of the range is quite abrupt, forming a part of the eastern rim of Jackson Hole, but at its eastern extremity a gentle dip slope leads down into the valley of Green River, where the beds reverse their dip to rise in the western flank of the Wind River Range. The northern boundary of the Gros Ventre Range is ill defined but has been arbitrarily placed at the Gros Ventre River. A well-defined topographic boundary exists along the south side of the range from its eastern termination to a point approximately 27 miles to the northwest,

but here the Gros Ventre and Hoback ranges join, making a separation on the basis of topography difficult. Here the boundary is better determined on the basis of structure. Most of the high peaks of the Gros Ventre Range reach elevations of approximately 11,000 feet.

The Hoback Range is a narrow, linear unit extending southward from the Gros Ventre Mountains for about 30 miles. It varies in width from but 4 to 6 miles. The peaks at its northern end have an average height of about 10,000 feet, but south of Hoback Canyon the range rises gradually, with Hoback Peak (10,818 feet) the culminating point.

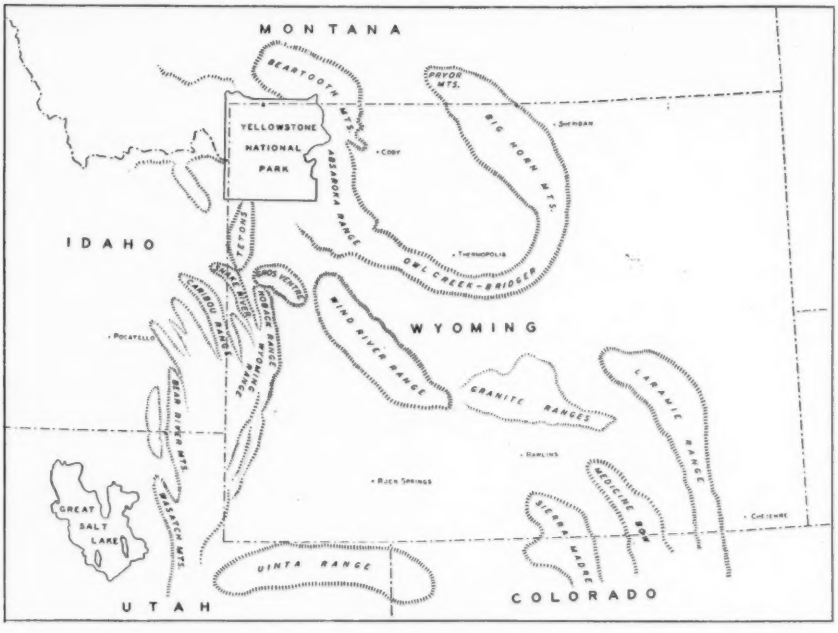


FIG. 1.—Index map

STRATIGRAPHY

The stratigraphic units used in mapping the area under discussion are listed in the chart (Fig. 2). Except for slight modification, they are the same as those used by Leland Horberg[1] in mapping the Teton Pass area to the west across Jackson Hole. The section as set forth in Figure 2 is a composite one, utilizing some of the formational names established by workers in the southeastern Idaho–southwestern Wyoming area, as well as some in use in northwestern Wyoming. The necessity of using such a composite section will be readily seen upon examination of the isopach map (Fig. 3) of portions of Wyoming, Idaho, and Utah. This map shows that the Rocky Mountain geosyncline had an irregular development. Locally there were basin-like areas where the pre-Laramide sediments accumulated in greater thicknesses than elsewhere. One such basin, or localized area of thick accumulation, centered

[1] "The Structural Geology and Physiography of the Teton Pass Area, Wyoming," *Augustana Library Pub. 16* (1938).

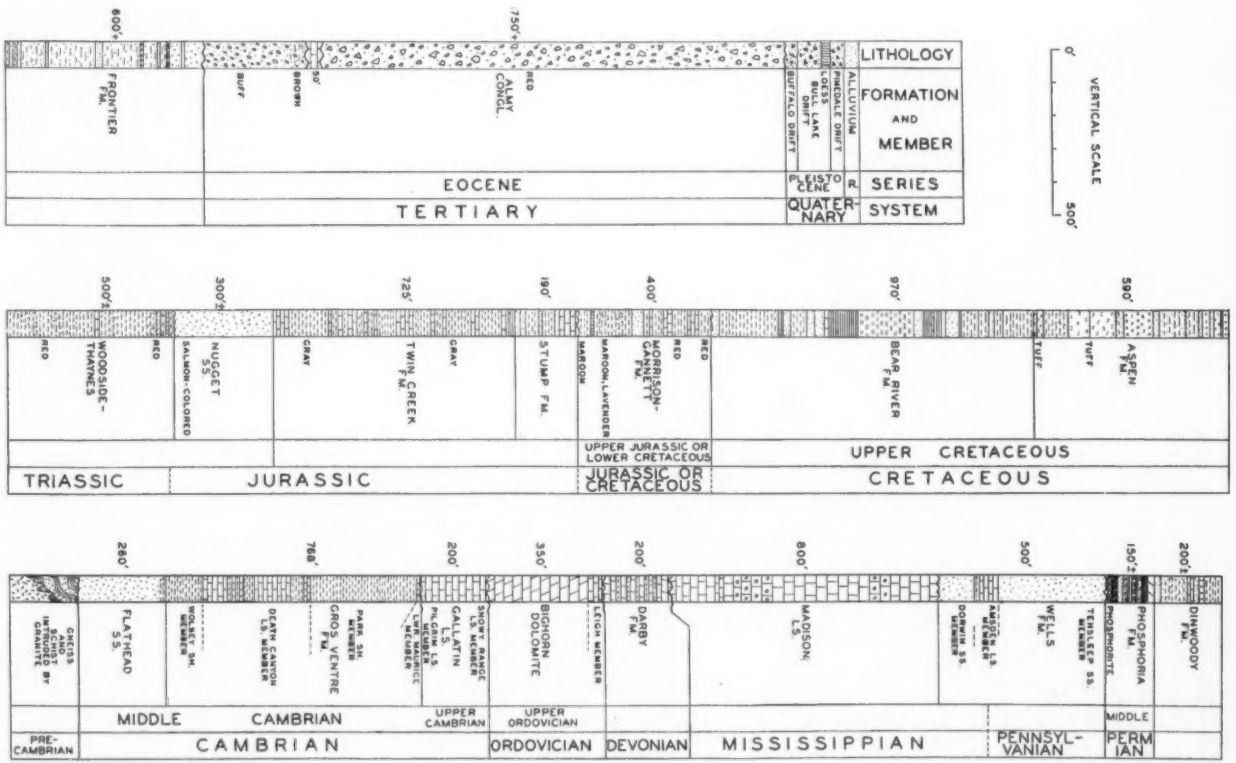


FIG. 2.—Stratigraphic chart

around the corner of Idaho, Wyoming, and Utah, where the Paleozoic and Mesozoic sediments attain a total thickness of over 43,000 feet. To the east and northeast the beds thin rapidly, indicating the existence of a foreland or platform area in that vicinity during the development of the geosyncline.

The Gros Ventre–northern Hoback area is located on the boundary between the foreland and geosynclinal tracts or in a region where most formations thicken rapidly to the southwest and pinch out to the northeast. However, in the later stages of its development or near the beginning of the Laramide orogeny the geosyncline enlarged greatly, spreading east and northeast. Thus the Upper Cretaceous sediments well within the foreland tract approach more closely the thicknesses of the beds of that age within the geosyncline proper than do the beds of the earlier periods.

Numerous stratigraphic problems exist in this boundary area, but the writers were primarily interested in the structural relationships, and therefore established formational names were applied to the mappable units.

The greatest difficulty in utilizing already established formational names was encountered in the Triassic sequence. The Triassic beds are here referred to the Dinwoody and the Woodside-Thaynes formations. The Dinwoody was named by Eliot Blackwelder[2] from the canyon of Dinwoody Lakes in the Wind River

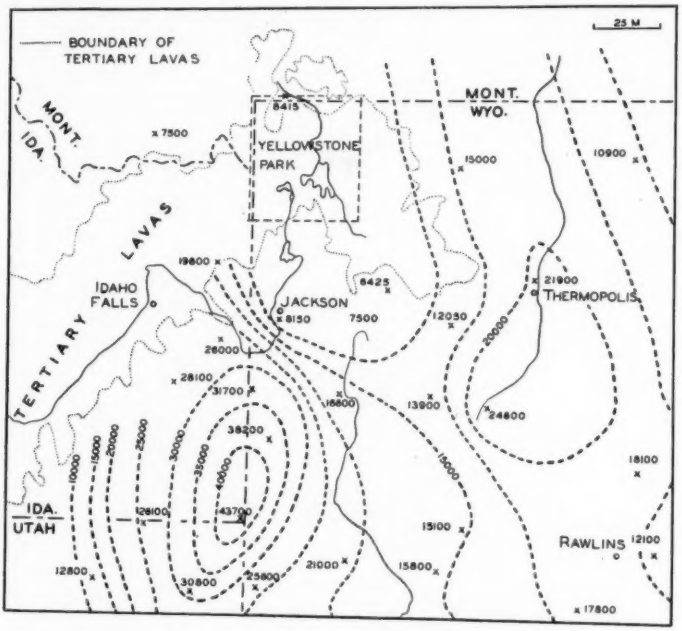


FIG. 3.—Isopach map (after Leland Horberg)

[2] "New Geological Formations in Western Wyoming," *Jour. Wash. Acad. Sci.*, Vol. VIII (1915), p. 425.

Range. Regarding its equivalents to the south and west, he stated:

> Westward it becomes progressively thicker, more calcareous, and more fossiliferous, and changes by imperceptible gradations horizontally into the Woodside and Thaynes formations of southeastern Idaho.

But more recent studies of the Triassic of western Wyoming by Norman D. Newell and Bernhard Kummel[3] have shown that the Dinwoody is the equivalent of only the lower portion of the Woodside of southeastern Idaho. While the mapping for this report was being done, the term "Ankareh" was used for the remaining Triassic above the Dinwoody. Newell and Kummel have found that this sequence of beds is equivalent faunally to the upper Woodside and Thaynes formations of southeastern Idaho. Therefore it now seems best to rename the beds that were once called "Ankareh" the "Woodside-Thaynes," using the hyphenated term.

REGIONAL TECTONIC SETTING

The controlling structures of the ranges in the Middle Rocky Mountain province in Wyoming, eastern Idaho, southern Montana, and northern Utah may be grouped readily into two distinct classes: foreland and geosynclinal.[4] The broad area in which the foreland structures are characteristic lies to the east of a narrower and more elongate belt of geosynclinal structures (see index map, Fig. 1). It should be noted that the pronounced structures of the foreland and geosynclinal areas do not approach one another closely except at the southern end of the Teton Range and at the junction of the Gros Ventre and Hoback ranges. These areas, therefore, are particularly significant, since the juxtaposition of these two fundamental types of structures should throw additional light on the mechanics of their formation and yield further information concerning the Laramide Revolution. Horberg has already described the structures of the Teton Pass area at the southern end of the Teton Range,[5] and a summary of the major structural types and trends of the larger Teton–Gros Ventre–Hoback area has been prepared in manuscript form.[6]

FORELAND STRUCTURAL TYPES

The foreland, or platform, structural types include high-angle faults (either normal or reverse), monoclinal flexures, domelike upwarps, and broad, open asymmetrical folds. These developed in response to stresses acting on regions in which the cover of Paleozoic and Mesozoic sediments was relatively thin (20,000 feet or less; see isopach map, Fig. 3). In such areas the structures of the underlying pre-Cambrian crystalline rocks have played the dominant role in determining the nature of the yielding to Laramide stresses.[7] Thus the general

[3] "Lower Eo-Triassic Stratigraphy, Western Wyoming and Southeastern Idaho," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), pp. 937–96.

[4] The concept of the two distinct types of structures, namely, foreland and geosynclinal, with reference to the Rocky Mountains has been developed by Dr. R. T. Chamberlin over a number of years in his classes and in field work in this region, making it impossible to cite an original reference in the literature.

[5] *Op. cit.*

[6] Leland Horberg, Vincent Nelson, and Victor Church, "Structural Types and Trends in Western Wyoming" (unpublished manuscript).

[7] W. T. Thom, Jr., "The Relation of Deep-seated Faults to the Surface Structures of Central Montana," *Bull. Amer. Assoc. Pet. Geol.*, Vol. VII (1923), pp. 1–13; Ernst and Hans Cloos, "Pre-Cambrian Structure of the Beartooth, the Bighorn and the Black Hills Uplifts, and Its Coincidence with Tertiary Uplifting," *Proc. Geol. Soc. Amer., 1933* (1934), p. 56; E. C. H. Lammers, "Pre-Cambrian Control of Laramide Structures in the Beartooth Range, Mon-

pattern of the platform region consists of a series of blocklike uplifts, many of which are distinctly rectangular in outline. The larger blocks or ranges are commonly composite in character, in that they usually consist of several small-

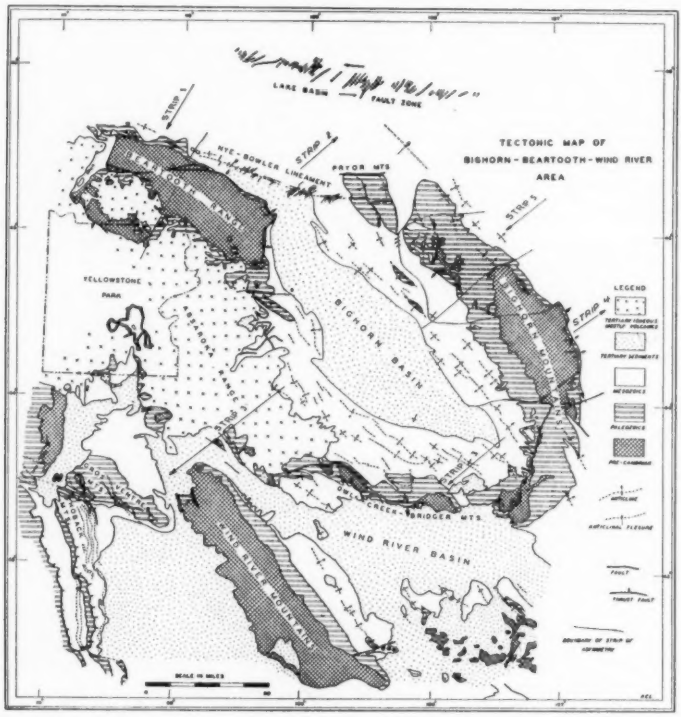


FIG. 4.—Regional tectonic map of Bighorn–Beartooth–Wind River area (after R. T. Chamberlin)

er blocks. With regard to such smaller blocks R. T. Chamberlin states:

> Some of the larger of these mountain units consist of subunits, or segments, each of which deformed with a certain individuality of its own. Characteristic of this individuality is a one-direction asymmetry of structural features which persists throughout each individual segment, but (in both the Beartooth and Bighorn ranges) is in the direction opposite from that of the adjoining segment on either side. The reversal of direction of structural asymmetry is one of the outstanding peculiarities of the tectonic development.[8]

The portion of Middle Rocky Mountain province of southern Montana, western Wyoming, and northeastern Utah which belongs to the foreland area includes the Pryor-Bighorn, Owl Creek-Bridger, Absaroka-Washaki, Beartooth, Teton, Gros Ventre, Wind River, and Uinta ranges. They all display clear-cut foreland structures, with the exception of the Teton Range, whose major features have resulted from Tertiary faulting. Two distinct groups of tectonic features

tana and Wyoming," *Bull. Geol. Soc. Amer.*, Vol. L (1939), p. 1918; D. L. Blackstone, Jr., "Structure of the Pryor Mountains, Montana," *Jour. Geol.*, Vol. XLVIII (1940), p. 606.

[8] "Diastrophic Behavior around the Bighorn Basin," *Jour. Geol.*, Vol. XLVIII (1940), p. 678.

are noted: (1) the dominating and controlling structures of the individual ranges, outlining or partially outlining the mountain blocks and (2) the secondary or minor structures superposed on both mountain blocks and basins alike. Chronologically, the initial or early phase of the Laramide deformation produced the major structural features and thus blocked out the mountain ranges. Later more surficial Laramide movements superposed secondary features on the major structures.

The combination of these primary and secondary foreland structures forms a unique and striking tectonic pattern. The axes of the secondary elements are all nearly parallel, having a strike of N. 40°–50° W. This strike is retained regardless of the trend of the ranges or basins of which they are a part, that is, irrespective of the strike of the major structural elements. Thus, for example, although the Beartooth–Absaroka-Owl Creek–Bridger–Bighorn–Pryor uplift forms a huge loop in which the major structural trends swing through 180°, the axes of the secondary elements remain constant at N. 40°–50° W. from one end of the loop to the other. Furthermore, although the secondary features are all approximately parallel and all asymmetrical, they are not all asymmetrical in the same direction. The foreland region seems to be divided into five more or less distinct strips or belts trending roughly northeast-southwest across the entire area, that is, across mountain ranges and basins alike. Within each of these strips the secondary features are all asymmetrical in the same direction (see tectonic map of Bighorn–Beartooth–Wind River area, Fig 4).

GEOSYNCLINAL STRUCTURAL TYPES

The structures of the narrow and elongate geosynclinal belt to the southwest of the foreland are in striking contrast to the tectonic features of the platform area. They have developed in response to stresses acting on a region where the sediments had attained great thicknesses (20,000–45,000 feet; see Fig. 3). These structures have been well summarized by Horberg as follows:

> Forming a relatively narrow belt, which follows either a northerly or northwesterly direction, the group is made up of a series of parallel structures which persist for long distances. The individual folds are closely compressed and often overturned to the east. Taken together they form a folded system of the Appalachian type. Probably the outstanding feature of the belt is a series of faults which to the north are exemplified by the Lewis, Lombard, and Heart Mountain faults, and farther south by the Darby, Absaroka, Bannock, Willard and Crawford faults. The horizontal displacement along some of these thrusts is probably more than 30 miles. Westward the structures merge with folding of Jurassic age.[9]

At least two other major low-angle thrusts should be added to those of the southern group mentioned by Horberg: (1) the St. John thrust in eastern Idaho[10] and (2) the Jackson thrust, first mapped and described by Horberg.[11]

In addition to the Jackson thrust, whose southwestward continuation into the northern Hoback region has been

[9] *Op. cit.*, p. 25.

[10] The St. John fault was first mapped by A. R. Schultz, "A Geologic Reconnaissance for Phosphate and Coal in Southeastern Idaho and Western Wyoming," *U.S. Geol. Surv. Bull. 680* (1918), p. 36. The name "St. John" fault was applied by Virgil R. D. Kirkham, "Geology and Oil Possibilities of Bingham, Bonneyville, and Caribou Counties, Idaho," *Idaho Bur. Mines and Geol. Bull. 8* (1924), pp. 33–34.

[11] *Op. cit.*, pp. 35–36.

mapped by the writers, at least three other low-angle, southwest-dipping thrusts have been encountered in the area mapped, namely, the Game Creek thrust, the Horse Creek thrust, and the Bear Creek thrust. These three thrusts, which lie south or southwest of the Jackson thrust, are both stratigraphically higher and smaller in extent. Thus the geosynclinal tract represents deformation of the surficial or thin-shelled type occurring where the basement crystallines were buried too deeply beneath weak sediments to be of much influence in shaping the structures, as contrasted to the thick-shelled type of the foreland region.

DESCRIPTION OF STRUCTURES

FORELAND STRUCTURES

The behavior of blocks typical of the foreland or platform area of the Middle Rocky Mountain province has been discussed by Chamberlin.[12] He noted that in cases where the steep sides of mountain blocks faced a basin the structures varied from simple flexures with high-angle faults to overturned anticlinal limbs and overthrust faults in various stages of development.

FIG. 5.—Cache Creek thrust fault between heads of Cache and Horse creeks. Inked line indicates trace of thrust plane on surface. Steep slope above thrust plane made up of (from botton to top) Mississippian, Devonian, and Ordovician beds. Smooth, gentle slope below thrust plane developed on Upper Cretaceous sediments. Sharp limestone ridge in foreground marks the northern edge of the Jackson thrust sheet near its abrupt eastern termination as shown on the right.

The more important structures of the Gros Ventre Range belong to this group; and the characteristic feature of the range is its blocklike pattern, particularly along

[12] *Op. cit.*, pp. 689–90.

the south flank where, throughout most of its extent, the steep mountainside faces Hoback Basin. Previous investigators described the range structurally as involving two parallel, northwest-trending anticlines, with axes about 5 miles apart.[13] The present study indicates that these anticlines are expressions of but two of the four or more foreland-type blocks which, with their related structures, form the basic tectonic pattern of the range. The blocks, three or four of which are distinctive enough to be termed "trap-door" structures,[14] are arranged serially along a N. 50° W. axis and from west to east are successively younger and less intense in degree of deformation. Beginning at the west, these blocks are (*a*) the Cache Creek thrust block, (*b*) the Skyline Trail block, (*c*) the Shoal Creek block with its associated structures, and (*d*) the Elbow Mountain block. In addition to these, other important structures of similar type, lying slightly to the north of the structural and topographic axis of the range, have been mapped. These include the Jackson Peak fault and the Nowlin Creek anticline.

[13] Orestes St. John, "Report of the Geological Field Work of the Teton Division," *Eleventh Ann. Rept. U.S. Geol. and Geog. Surv. of Territories, Embracing Idaho and Wyoming, 1879*, pp. 448–56.

[14] A descriptive term sometimes applied to a blocklike type of foreland structure, the essential feature of which is a sharply asymmetrical flexure usually broken on its steep flank by a high-angle fault. The flexure or fault or both generally outline two adjacent sides of the block, and displacement is commonly greatest at the corner. The fault surface curves in under the block, and in more advanced stages actually becomes a thrust fault, indicating its ramplike character. Trap doors are thus interpreted as resulting largely from horizontal stresses transmitted through the more competent crystalline complex below the sedimentary cover to pre-existing zones of weakness, where they break up toward or through to the surface to produce a fault having a "sled-runner"–like profile.

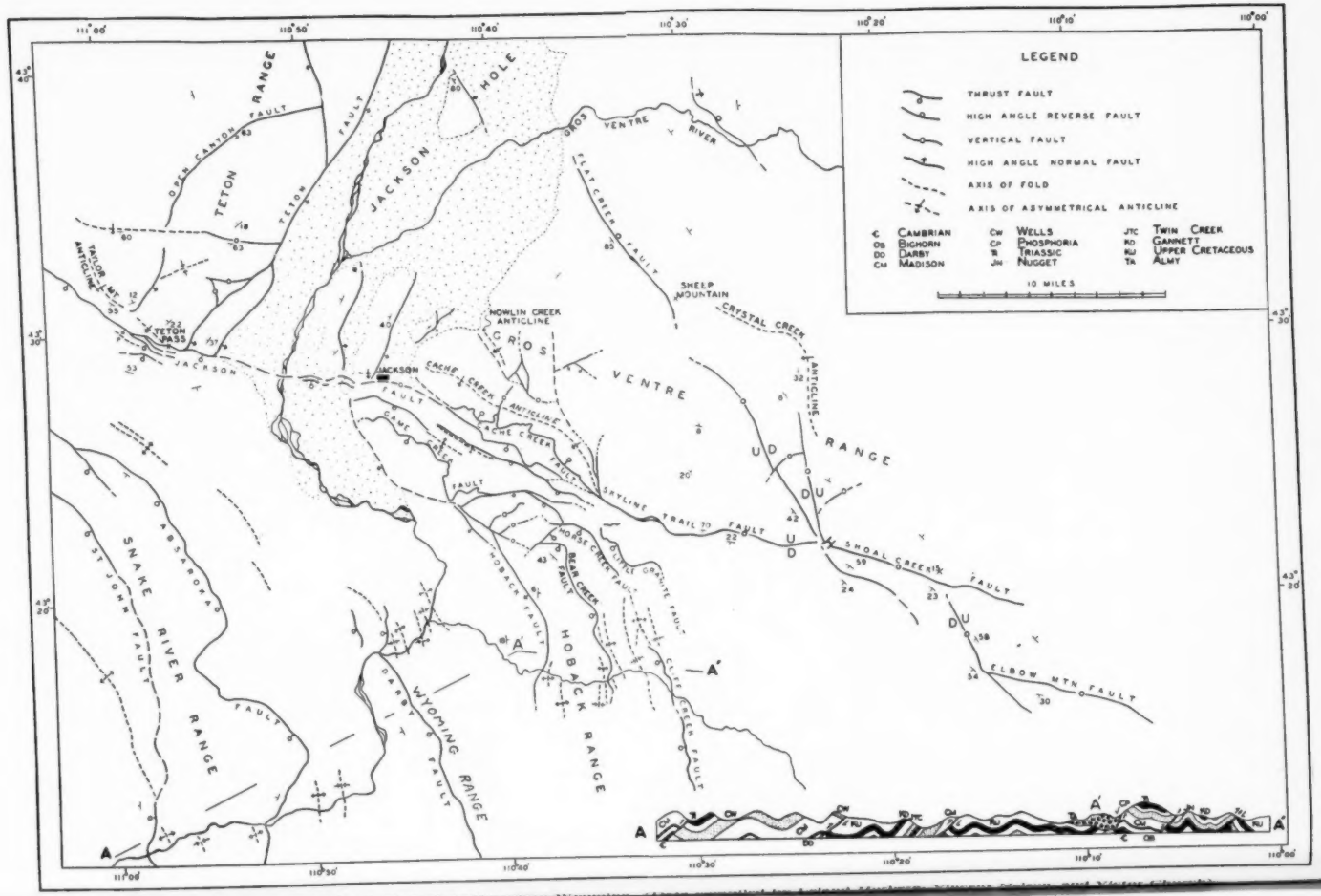
CACHE CREEK THRUST BLOCK

The dominant tectonic feature of the west end of the Gros Ventre Range is the Cache Creek anticline and the associated Cache Creek thrust fault.

The Cache Creek anticline is a broad, asymmetrical flexure, which can be traced for about 9 miles striking N. 45° to N. 50° W. It is essentially monoclinal in character, having a steep south limb (Figs. 6 and 8, Sec. *A*–*A'*) which parallels Cache Creek. The dip of the beds on this steep south limb varies from 60° to the south to 50° to the north, where overturn has been strongest. The structure terminates abruptly at the eastern margin of Jackson Hole, either by cross folding of synclinal character or by faulting. Horberg[15] has suggested that the Cache Creek anticline may once have been continuous across Jackson Hole and that its western continuation may be found in the Teton Range as the Taylor Mountain anticline. Eastward the structure loses its distinctiveness and terminates against the Skyline Trail block.

The Cache Creek thrust, representing the broken southwest limb of the Cache Creek anticline, can be traced from the Skyline Trail block westward to Jackson Hole—a distance of approximately 10 miles. Throughout this distance the fault is easily traceable on the north wall of Cache Creek Valley (Fig. 5). Beds exposed on the front of the thrust plate range in age from Jurassic (Twin Creek) to Ordovician (Bighorn), having the steep dips already mentioned in connection with the Cache Creek anticline. These beds have been thrust over Upper Cretaceous (Bear River?) rocks which show strong overturn, dipping 75°–40° N. Nowhere along the trace of the thrust is it possible to observe the fault surface

[15] *Op. cit.*, p. 33.


LEGEND
THRUST FAULT
HIGH ANGLE REVERSE FAULT
VERTICAL FAULT
HIGH ANGLE NORMAL FAULT
AXIS OF FOLD
AXIS OF ASYMMETRICAL ANTICLINE
€ CAMBRIAN
OB BIGHORN
D DARBY
C M MADISON
C W WELLS
C P PHOSPHORIA
T R TRIASSIC
J N NUGGET
J T C TWIN CREEK
K D GANNETT
K U UPPER CRETACEOUS
T A ALMY
10 MILES
TETON RANGE
JACKSON HOLE
GROS VENTRE RIVER
SNAKE RIVER RANGE
WYOMING RANGE
HOBACK RANGE
GROS VENTRE RANGE
JACKSON
TETON PASS
SHELP MOUNTAIN
CRYSTAL CREEK
NOWLIN CREEK ANTICLINE
CACHE CREEK ANTICLINE
SKYLINE TRAIL FAULT
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itself, but its sinuous trace gives ample evidence that it dips to the northeast at an angle of probably no more than 30°–35°. Horizontal displacement is relatively small, probably nowhere exceeding 1 or 1½ miles. To the west the fault is covered with the alluvial material of the lower portion of Cache Creek and Jackson Hole, while to the east it terminates against the Skyline Trail fault. Only one small tear fault, having a N. 30° E. strike, cuts the Cache Creek thrust. This tear fault produces a small offset in the thrust about 5 miles east of Jackson.

The Cache Creek thrust block, including the Cache Creek anticline, is the earliest formed and most strongly developed of the series of foreland blocks making up the Gros Ventre Range. The forces producing this block were similar to those producing the ramplike trap doors to the east but operated here with such great intensity that accommodation had to be effected by thrusting rather than by high-angle faulting.

SKYLINE TRAIL BLOCK

The second of the foreland blocks is outlined by the Skyline Trail fault (Fig. 8, Sec. *C*–*C*′), a vertical or high-angle reverse fracture extending for almost 10 miles along a nearly east-west strike before curving to the north near its western end. After turning north it dies out rapidly, disappearing in about 2 miles. At its eastern extremity the structure is cut by the Shoal Creek fault of the third foreland block. Displacement along the fault reaches a maximum at the southwest corner or apex, where the structure swings around to the north. At this point the stratigraphic displacement can be hardly less than 6,000 feet and may be 8,000 feet or more. Precise determination of displacement is not possible, for pre-Cambrian rocks are in apparent contact with the Almy (Tertiary) formation, which was deposited after at least the first phase of the Laramide deformation.

Following the Skyline Trail fault eastward from the apex of the block, displacement diminishes regularly at about 350–400 feet per mile. Just west of its termination against the Shoal Creek fault the Skyline Trail fault has a minimum stratigraphic displacement of about 1,000–1,200 feet (Madison to Dinwoody). Actual displacement may be somewhat greater, however, since both formations are essentially vertical.

North of the fault the Paleozoic beds dip evenly to the northeast at angles ranging from 10° to 20° (Fig. 9). Near the fault the beds flex over sharply, assuming a high southwest dip or even becoming vertical in some places. The Mesozoic formations in contact with the Paleozoics at the fault generally have similar attitudes, but at some points they are even overturned.

Closely associated with this structure is a high-angle fracture, the Sheep Creek fault (Fig. 7), which branches off on the west side of the north extension of the Skyline Trail fault. For a mile or so it diverges but little from the Skyline Trail fault, but then it swings to the northwest for 1 or 2 miles before returning again to a north-south strike. This structure is of particular interest, since displacement along it increases from almost nothing at its juncture with the Skyline Trail fault to 700 or 800 feet directly east of Jackson Peak and is in reverse relation to that of the Skyline Trail fault, that is, the west side is the upthrow side. This peculiar pattern, suggesting modified pivot or scissors faulting, assumes added significance, since almost identical relations are shown in connection with the Shoal Creek–Pyramid Peak faults to the east.

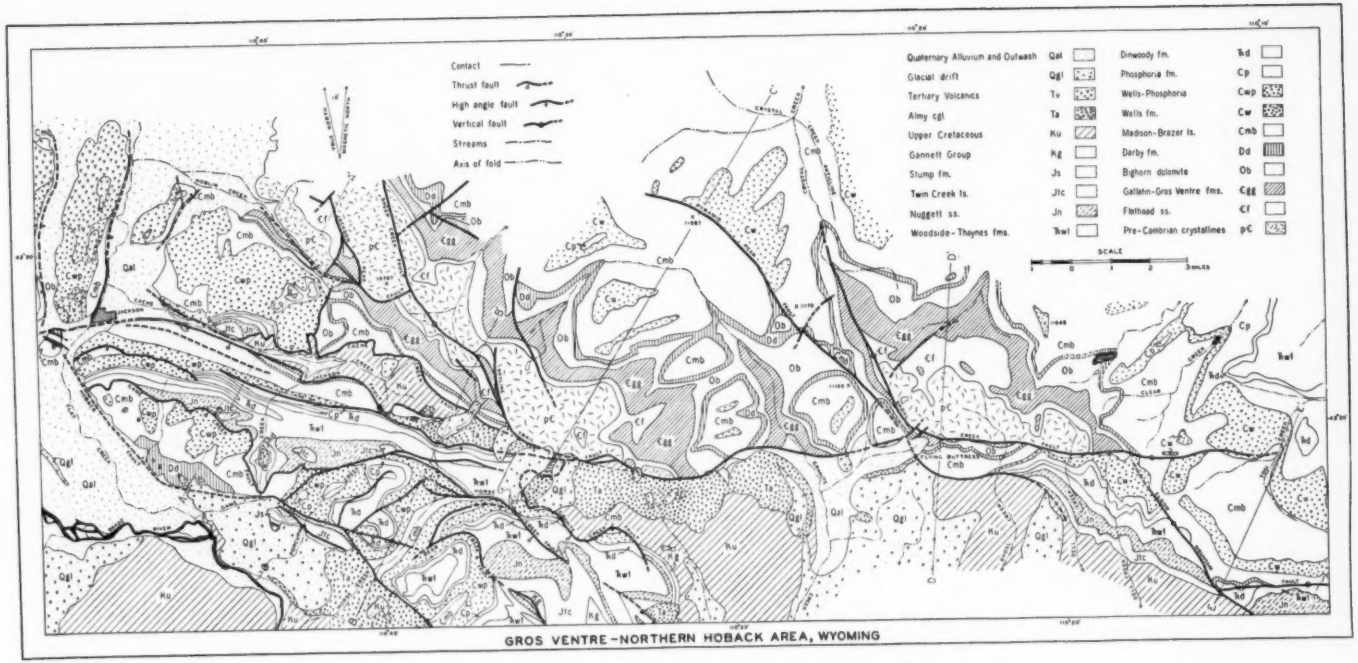

FIG. 7.—Areal geology of the Gros Ventre–northern Hoback area

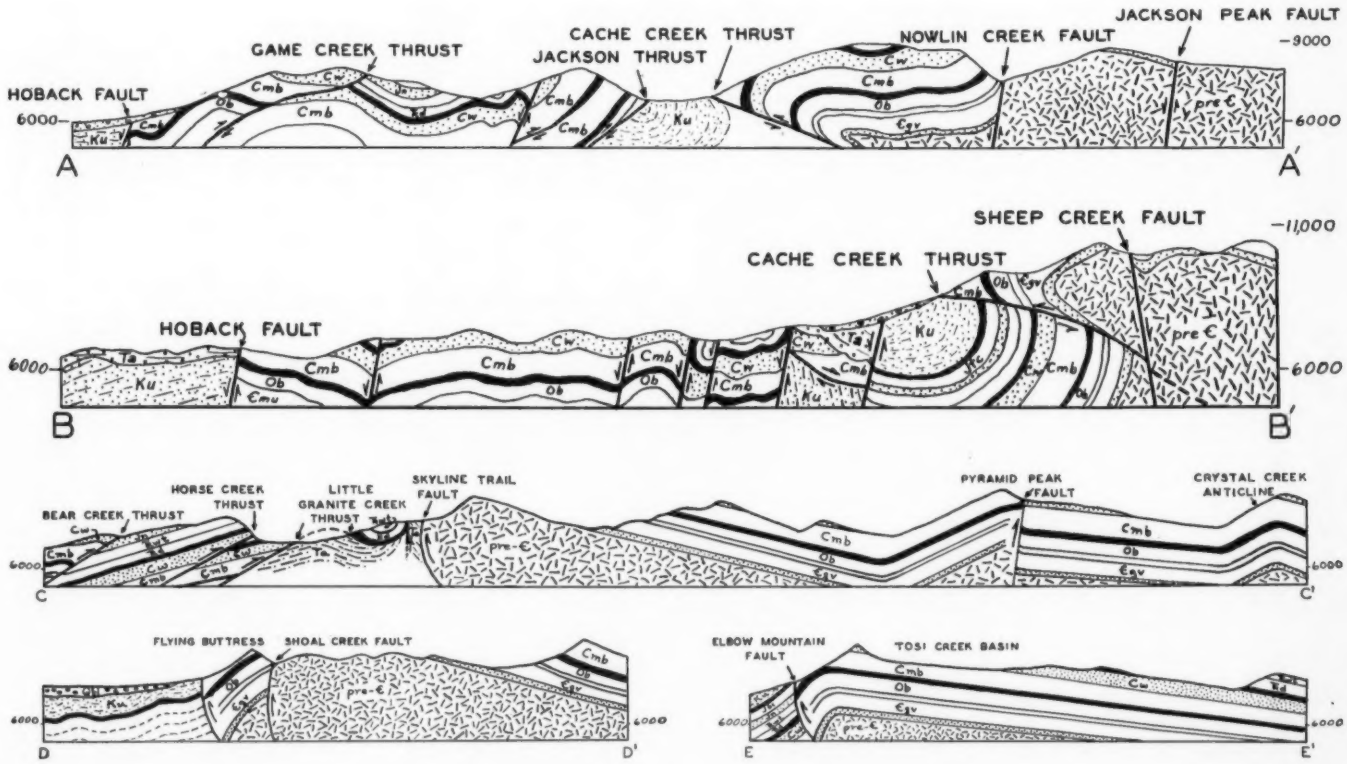

FIG. 8.—Cross-sections of the Gros Ventre–northern Hoback area

SHOAL CREEK BLOCK AND ASSOCIATED STRUCTURES

The Skyline Trail structure terminates abruptly against the Shoal Creek fault, a conspicuous fracture which bounds the western and southern sides of the next well-defined trap-door block. It is an essentially vertical or high-angle normal fracture (Fig. 8, Sec. *B–B′*), with an average strike of N. 75° W. along the southern side of the block. The strike swings rapidly to N. 10° W. just west of the Shoal–Swift Creek drainage divide and continues northward for about 3 miles before dying out. Along the southern side of the block the fault may be traced for nearly 9 miles, when it gradually disappears within the Madison formation in Tosi Creek Basin. Like the Skyline Trail fault, the displacement is greatest along the southern margin of the block. Maximum displacement, which occurs about 2½ miles east of the southwest corner, is 2,000–3,000 feet (pre-Cambrian to Wells). At the southwest corner of the block, where a sharp notch (elevation 9,500 feet) has developed between two 11,000-foot peaks, the displacement is only 1,100 feet (lower Flathead to upper Gallatin).

South of the Shoal Creek fault near its western end, and almost detached by it from the rest of the range, is a high, prominent peak (11,100 feet) called the "Flying Buttress" (Fig. 10). The strata on the south side of this tectonically significant peak attain dips of 65°–75° to the south. On the north side the beds are cut off sharply to form an abrupt massive wall of remarkable linearity. This north face of the Flying Buttress has been developed along the Shoal Creek fault, while on its south and west sides it is outlined by a fault which swings

FIG. 9.—Gently dipping Paleozoic beds on north slope of the Skyline Trail block

around in a sweeping arc quite similar to the changes in strike seen along the Skyline Trail, Shoal Creek, and Elbow Mountain faults. Maximum displacement is at least 2,500 feet and may be even more (Madison to Upper Cretaceous). Since this fault has been cut by the Shoal Creek fault, it is evident that it developed during the first phase of the foreland deformation of the Laramide orogeny. The Shoal Creek fault represents a second phase. Thus the Flying Buttress is a nose or slice of the south side of an original trap-door block left behind when the second phase of deformation produced a fault which broke through a little to the north of the original fracture.

The Crystal Creek anticline (Fig. 7) represents the northward continuation of the Shoal Creek structure. The anticline is a broad, open structure, asymmetrical to the southwest with dips as high as 35° on the steep flank. The axial high of the structure lies to the east of the upper part of Crystal Creek. It turns west along the west branch of Crystal Creek before swinging north and dying out on the north flank of the range east of Sheep Mountain.

Near the southwest corner of the Shoal Creek fault a minor vertical fracture—the Pyramid Peak fault (Fig. 7)—developed and strikes northwest across the central part of the range. The displacement is slight near the Shoal Creek fault but increases steadily to the northwest, until at Pyramid Peak it is estimated to be 1,000–1,200 feet. The upthrow side of this fault is on the west, or reversed from that on the Shoal Creek fault. Thus the relations are strikingly similar to those existing between the Sheep Creek and the Skyline Trail faults, indicating that

FIG. 10.—Steeply dipping strata on the south side of the Flying Buttress as viewed from the west

both the Pyramid Peak and the Sheep Creek faults are closely related in origin.

ELBOW MOUNTAIN BLOCK

South and east of the Shoal Creek block lies the block outlined by the Elbow Mountain fault (Fig. 8, Sec. E–E'). The name for this fault, derived from the prominent mass forming most of the southeast portion of the Gros Ventre Range, is singularly descriptive of the structural, as well as the topographic, features of the block. Elbow Mountain is an elevated mass of Paleozoic sediments, cut off on the south and west to form a bold escarpment. The bordering fault of this block can be traced for 5 miles along the south side and 3 miles along the west side of Elbow Mountain. Its strike is more regular than that of the Cache Creek thrust, Skyline Trail, or Shoal Creek faults, and averages N. 80° W. on the south side and N. 30° W. on the west. The displacement, considerably less than along any of the previously discussed major block structures, reaches a maximum at the southwest corner, where it is about 2,500 feet (Bighorn to Twin Creek). The amount of movement on the fault diminishes along the strike both to the north and to the east, and the fault dies out in the Wells formation at both extremities. To the north there is no folding to represent the continuation of the structure, but to the east a monoclinal flexure along Rock Creek apparently continues the displacement in decreasing amount to the east end of the range.

A fault of minor importance—the Jack Creek fault—branches from the Elbow Mountain fault at the corner of the block and strikes southeast. Displacement is moderate (1,000 feet or less) near Elbow Mountain and diminishes southeast. South of Jack Creek the fault is lost under the cover of Tertiary conglomerate.

JACKSON PEAK FAULT

This structure is a high-angle or vertical fault which outlines the south and west sides of the pre-Cambrian block, dominated by Jackson Peak in the northwest portion of the range (Fig. 7). The Jackson Peak fault terminates either by intersection with the Sheep Creek fault about ¾ mile southeast of Jackson Peak or dies out very rapidly within the outcrop area of pre-Cambrian rocks in that vicinity. From its eastern extremity the fault strikes westward for approximately 1 mile and then curves sharply to the north, following the drainage line of a small stream to the northern boundary of the mapped area. Maximum displacement along the structure occurs at its southwest corner (pre-Cambrian to Gros Ventre) and is difficult to determine but is estimated at no less than 2,200 feet and it may be even more. The block thus outlined by the Jackson Peak fault resembles closely the typical foreland-type blocks to the east and probably should be regarded as a trap-door block.

NOWLIN CREEK ANTICLINE

This structure lies about 3–4 miles north of and parallel to the Cache Creek anticline (Fig. 6), which it resembles in many respects. It is asymmetrical to the southwest, terminates abruptly to the west at Jackson Hole, and disappears to the east near the Skyline Trail block. Here the Nowlin Creek and Cache Creek flexures apparently merge. Horberg[16] has suggested that the Nowlin Creek anticline may continue westward across Jackson Hole and reappear in the Teton Range as the Moose Creek anticline.

[16] *Ibid.*

GEOSYNCLINAL STRUCTURES

THRUST FAULTS OF THE GEOSYNCLINAL BELT

The Jackson and Little Granite thrusts.—In the western portion of the area under consideration the northern margin of the geosynclinal tract is marked by the southwest-dipping Jackson thrust, which was first mapped by Horberg.[17] In his area the fault may be seen emerging from the Tertiary lava cover just a little east of Victor, Idaho, thence continues east through Teton Pass across Jackson Hole and into the northern edge of the Hoback Range. Here it may be traced southeast from the east margin of Jackson Hole for a distance of about 8½ miles (Fig. 7), having an average strike of about S. 60° E. The thrust follows along the south wall of Cache Creek Valley, giving Cache Creek a remarkable structural setting, since it occupies a position between two thrusts, the northeast-dipping Cache Creek thrust on the north and the southwest-dipping Jackson thrust on the south (Fig. 8, Sec. *A–A′*).

Nowhere along its trace is the plane of the Jackson thrust exposed; but, judging from the course of the fault, the plane must dip at a relatively low angle to the southwest. Along most of the fault, stratigraphic relations show nearly vertical Carboniferous (Madison ls.) resting on Upper Cretaceous beds. However, at one exposure approximately 4 miles up from the mouth of Cache Creek Valley, Upper Cambrian beds were observed at the front of the thrust sheet, suggesting that the thrust mass may possibly have ridden on Middle Cambrian shales.

Some faulting of the imbricate type is unquestionably associated with the Jackson fault. The Madison limestone found at the front of the thrust sheet is far too thick to represent normal relationships. Poor exposures on the entire north-facing slope of Cache Creek Valley made it difficult to trace such slice faults, but evidence of such faulting was found at one place where the Darby limestone occurs between outcrops of almost vertical Mississippian (Madison) limestone. Though the exact course of this minor fault could not be traced for any distance, such a fault has been indicated on the map (Fig. 7), and it is quite possible that several other small slices are present within the Madison limestone.

The Jackson thrust sheet terminates abruptly as a nose of vertical Madison limestone in contact with Almy (Eocene?) conglomerate about 8½ miles southeast of Jackson Hole.

About 3¼ miles southeast of the eastern termination of the Jackson thrust a second thrust mass from the southwest is encountered. This has been called the "Little Granite" thrust, from exposures along the creek of that name. The Little Granite thrust is exposed first in the eastern part of the Horse Creek drainage area (Fig. 7), crosses the Horse Creek–Little Granite drainage divide, and continues east for a short distance along the southern side of the upper part of Little Granite Creek before swinging south and dying out in a fold. The general relations along the fault show late Paleozoic and early Mesozoic beds thrust over Cretaceous formations, but in one exposure in a small basin-shaped valley near the western extremity of the thrust extremely significant relations are shown (Fig. 8, Sec. *C–C′*). Here late Paleozoics, making up a portion of the thrust sheet, are resting on the Almy conglomerate of early Tertiary age. The exposure of the Almy below the thrust sheet is a window in the overriding block, for the thrust relations seen on the south side of the basin are repeated on the north side ¾–1 mile

[17] *Ibid.*, p. 35.

away. Patches of till on both the east and the west sides make it impossible to trace completely the outline of the window. The fault plane on the north side of the window dips to the north or in the opposite direction from the normal dip of the fault. This change in dip seems to be due to later faulting and associated folding which affected the region sometime after the major Laramide deformation.

Mention has already been made of the sudden termination of the Jackson fault at its southeast end and the difficulties encountered in attempting to trace the Little Granite thrust westward from the window of Almy because of glacial-drift cover. There are then two possible interpretations for the relations between these faults: (1) they are two separate thrusts; (2) they are one continuous thrust.

At the east end of the Jackson thrust vertical beds of Madison limestone are in contact with the Almy conglomerate (Eocene?). Just a little west of this point the Almy occurs at higher elevations than the thrust plate and also covers small portions of its front. Thus in this area the age of the Almy is post-thrusting. No doubt the material composing the Almy conglomerate came from the area of the high Gros Ventres lying to the north, where the Mesozoic beds have been stripped away. In many places the material of the Almy is extremely coarse, so that originally the formation must have had a moderate initial dip to the south; but the Almy, as it now lies adjacent to and east of the Jackson thrust, dips steeply to the north (45°). Traced eastward, the formation gradually flattens out and finally retains only the original south dip.

A normal fault lies a short distance south of and roughly parallel to the Jackson thrust fault but continues southeastward beyond the easternmost exposure of the Jackson thrust. The reversal of dip of the Almy conglomerate directly east of the termination of the Jackson thrust has led the writers to suspect a second high-angle, arcuate, hinge-like fault that branches from the first-mentioned fault and strikes N. 65° E. to east-west. This fault cut across the Jackson thrust sheet and the overlying Almy, producing a hinge block between it and the normal fault to the south. The north side of this hinge block was dropped, burying the southeastern part of the Jackson thrust sheet and tilting the Almy to the northeast (Fig. 8, Sec. *B–B′*). Since the fault dies out to the east, as indicated by the gradual flattening of the Almy, the Jackson fault may reappear at the surface farther on as the Little Granite thrust. If the Jackson and the Little Granite thrusts are one and the same, it is necessary to assume that there was renewed movement on the southeastern portion of the thrust in post-Almy time, since the Almy is exposed in the window at the western end of the Little Granite fault. Despite the rather specific set of conditions required by the hypothesis that the Jackson and Little Granite thrusts are one single fault, the writers favor this hypothesis for the following reasons: (1) the two faults are separated from each other by a distance of less than 3 miles; (2) the faults are striking in directions which should bring them together; and (3) they are the outermost thrusts in a relatively wide zone of thrusting.

The Game Creek thrust.—In the western portion of the area the front of a second thrust sheet is encountered about 2½ miles southwest of the Jackson fault (Fig 8, Sec. *A–A′*). It has been called the "Game Creek" thrust, for the plane of the fault is well exposed on the valley

sides of Game Creek, which forms a large re-entrant in the thrust sheet. Exposures in the valley are especially striking, since the gray Mississippian limestone has been thrust over the brilliant red and brown Triassic formations.

The Game Creek thrust may be traced from the eastern margin of Jackson Hole to the southeast for approximately 7 miles. Its trace roughly parallels the strike of the Jackson thrust. The front portion of the thrust plate is made up of Carboniferous beds which have been thrust over closely folded Triassic and Jurassic beds. From the exposures of the fault surface on the sides of Game Creek Valley it may be seen that the thrust plane is essentially horizontal for a distance of about 1 mile southwest from its front, but then it dips steeply to the southwest as a result of drag on a normal fault which cuts the thrust plate. South of this normal fault the plane of the thrust lies below the surface.

High-angle faulting has made it impossible to trace the Game Creek fault for more than the 7 miles already specified. The western margin of the Hoback Range is outlined by a high-angle fault (the Hoback fault), having its downthrow side to the west. At the northwest end of the range this fault produces truncation of all the northwest-trending structures, including the Game Creek fault, at the east margin of Jackson Hole. In following the trace of the Game Creek plane southeastward, one loses it on the divide between Game and Porcupine creeks. Here the thrust is again cut by a high-angle fault which strikes nearly east-west and has its downthrow side to the south, with the consequent dropping of the thrust plane below the surface.

The Horse Creek thrust.—The thrust relationships of this fault are clearly shown just a little east of the junction of the two main branches of Horse Creek. Here the plane of the fault is exposed at creek level, with nearly flat-lying Madison limestone resting on almost vertically standing beds of late Paleozoic and early Mesozoic age, which strike at nearly right angles to the trace of the fault plane. Traced to the east, southeast, and south, the fault follows closely the east branch of Horse Creek (Fig. 8, Sec. *C–C′*) before swinging south near the head of the creek. Along this course it cuts higher both topographically and stratigraphically. In all, the fault may be traced for a distance of about 4 miles. To the south it terminates in a fold within the Nugget formation, while westward the stratigraphic displacement decreases from the maximum of Madison over Woodside-Thaynes, near the head of the creek, to Madison over Wells and then dies out within the Wells between Horse and Porcupine creeks.

The Bear Creek and Cliff Creek thrusts.—Two other thrusts in the northern end of the Hoback Range have been traced by Horberg, who extended the writers' mapping southward across Hoback Canyon. The northern end of one of these—the Bear Creek thrust—is shown on the map (Fig. 8, Sec. *C–C′*). It follows the east border of the main crest of the range, trending N. 20° W., and, unlike the other structures, does not reflect the change of trend from north-south to northwest at the northern end of the range. The fault is clearly related to folding and appears to represent a broad anticlinal structure thrust eastward over a compressed syncline. Stratigraphic displacement is at no point over 1,000 feet. Northward the fault plane flattens and disappears along bedding planes in the Wells formation. The structure has been traced south to the Hoback River, where the displacement

is small, and normal fold relations appear to exist to the south.

Although only the northernmost end of the Cliff Creek thrust is within the map limits, its relation to the general tectonic pattern (see Fig. 6) is so important that it should be described briefly. The Cliff Creek thrust marks the eastern border of the Hoback Range. Along most of the thrust front the Nugget sandstone is in contact with the Evanston (Fort Union) formation, in which Hoback Basin to the east has been eroded. The outer margin of the thrust mass is represented by discontinuous monoclinal ridges and isolated noses developed along the strike of the Nugget sandstone. The northernmost hogback is found immediately east of the mouth of Little Granite Creek, and from this point similar features continue south as conspicuous land forms for a distance of 18 miles.[18] North of the northernmost hogback the fault dies out rapidly in a sharp fold. Although the Evanston beds east of the fault have been sharply folded in places, the Cliff Creek fault may properly be considered the eastern border of the geosynclinal deformation.

Relationships of the thrusts.—A notable feature of those ranges within the geosynclinal tract (the Hoback and Snake River ranges in particular) is the sharp swing in the strike of their main structures where they abut the foreland region (Gros Ventres and Tetons). Throughout most of the Hoback Range the structures trend north-south, but at its northern end they curve sharply to the northwest. The Cliff Creek fault, marking the eastern boundary of the range through most of its course, dies out at the northeast corner of the range. Here, at the point of curvature between the northern and eastern boundaries of the Hoback Mountains, there is a series of small, discontinuous thrusts taking the place of a single boundary structure. These thrusts have served to take up the stresses by slicing around the Gros Ventre buttress. Nearly all these small thrusts have northwest strikes at their northern ends and curve to the southeast and south at their opposite ends. Westward from the strong arc between the eastern and northern boundaries of the range, a single continuous fault (the Jackson thrust) again forms the main boundary structure of the geosynclinal region in both the western portion of the map area and the Teton area across Jackson Hole to the west. Of the several thrusts described, perhaps only the Jackson and the Cliff Creek structures should be considered major features. The other faults (i.e., the Bear Creek, Game Creek, Horse Creek, and perhaps the Little Granite, if it is not a part of the Jackson fault, as here supposed) collectively make up the continuation of the thrust zone represented by the Jackson fault on the northern, and the Cliff Creek fault on the eastern, boundaries of the curving geosynclinal tract.

Folding associated with the thrust faulting.—Prior to the development of the numerous thrust faults in the geosynclinal tract the rocks underwent severe deformation by folding. The stresses responsible for this folding were so great and the deformation progressed so far that finally the rocks were forced to yield by thrusting. That much of the tight folding preceded the major thrusting can be seen upon examination of the beds lying below the thrust planes. In many places the Triassic and Jurassic formations particularly have been thrown into a series of folds too close and numerous to be plotted on the map. Not un-

[18] See Gros Ventre Quadrangle, Wyoming, topographic map.

commonly, folding is localized within a single formation, and in several places chevron folds have developed on a comparatively large scale.

HIGH-ANGLE FAULTING

Several high-angle normal faults of Tertiary age are found in the northern Hoback Range. The most important of these is the Hoback fault (Figs. 6 and 7), which outlines the western margin of the Hoback Range. This fault has been accurately traced from a point about a mile south of Hoback Canyon to Game Creek. North of Game Creek the fault itself cannot be seen, but there is ample evidence to indicate its presence along the base of the range. The west-facing slope of the range has the appearance of a scarp, and, beyond the front of the Game Creek thrust sheet, nearly vertical west- and northwest-striking beds terminate abruptly to form the east rim of Jackson Hole. These same beds with similar strike reappear a short distance to the west in the small "butte" southwest of Jackson, indicating that they were faulted out of the intervening area. Relations along the Hoback fault between Game Creek and the ridge separating Porcupine and Horse creeks show Nugget resting against Wells. South of this the Almy conglomerate lies on the downthrow side adjacent to Paleozoic and Mesozoic beds.

Several faults of minor importance, having predominantly northeast strikes and terminating against the Hoback fault, lie in the area between Game and Horse creeks. One of these is the fault which cuts off the southeastern extension of the Game Creek thrust. This fault can be traced from its intersection with the Hoback fault to a point about 6 miles to the east before dying out in the Upper Woodside-Thaynes.

A relatively long normal fault lies a short distance south of and roughly parallel to the Jackson fault. It continues eastward beyond the easternmost exposure of the Jackson fault and in so doing has dropped the Phosphoria formation (a part of the Jackson thrust sheet) against the northeast-dipping Almy conglomerate, which here covers the front portion of the thrust mass.

DEVELOPMENT OF STRUCTURES

The descriptions of structures have shown that there are marked differences in the tectonic patterns of the geosynclinal and foreland tracts, but the sequence of development of these types must still be considered.

PRE-LARAMIDE DIASTROPHISM

The only pre-Laramide diastrophism of any importance to affect the area occurred in pre-Cambrian time. The crystalline rocks indicate long periods of sedimentation, folding, igneous activity, and erosion prior to the deposition of the Middle Cambrian sands. This pre-Laramide diastrophism is of importance in understanding the present structures, for it is believed that the structural trends and lines of weakness developed then were effective in determining the tectonic pattern assumed in the foreland area during the Laramide revolution.

DEVELOPMENT OF LARAMIDE STRUCTURES

The Laramide orogeny, which was responsible for the development of the open asymmetrical folds and trap-door structures of the Gros Ventre area, as well as the close folds and associated thrust faults of the Hoback region, began after the deposition of the Frontier formation. From evidence available within the Gros Ventre–Hoback area, the dating of the close of this period of deformation

presents somewhat more of a problem. Field evidence, such as that presented in the discussion of the trap-door blocks and Jackson thrust, indicates that most of the movement preceded the deposition of the Almy. However, some post-Almy movement is indicated in the Little Granite thrust, as well as on the Skyline Trail and Shoal Creek faults. It is not possible with the evidence available within the area to date this post-Almy movement, but it is considered of secondary importance and not the main episode of the Laramide deformation. The age of the Almy has not been fixed definitely; but beds of the Evanston formation, which lie stratigraphically below the Almy, have been sharply folded east of the Hoback Range in Hoback Canyon. The Evanston is thought to be the equivalent of the Fort Union, indicating that the deformation continued well into Paleocene time.

Thus the deformation responsible for the foreland and geosynclinal structures extended over a considerable period of time, and it seems likely that the two types of structures may not have developed simultaneously. The surficial movement produced in the geosynclinal and foreland areas by the causal forces was in opposite directions. This is evident from examination of nearly all the structures found in the two areas. In the foreland tract virtually all the major structures are asymmetrical to the southwest. The major folds, that is, the Cache Creek anticline, Nowlin Creek anticline, etc., have their steep limbs to the southwest. Thrusting on the Cache Creek thrust has been from the northeast; and the trap-door blocks have been interpreted as ramp structures, indicating some active horizontal stress from the northeast. In the geosynclinal area all the thrust-fault planes dip southwest or south with overriding toward the east and northeast.

It is possible that the Laramide stresses may have produced the surficial movements in opposite directions simultaneously, but a more likely interpretation would be that the causal forces produced surficial movement in the one area as long as that area could give the required crustal shortening most easily. Later the deformation could shift to the adjoining area, producing surficial movement which might very well be in an opposite direction. Evidence of such two-phase deformation is to be found in the area of the Jackson and Cache Creek thrusts. It seems reasonable to believe that the fronts of both the Jackson and the Cache Creek thrust masses have been stripped back by erosion considerable distances from their original positions. This means that, if the two thrusts developed simultaneously, their overlap would have been much greater than if one had developed earlier than the other and probably great enough that some remnants of the overlap should be found, but they are entirely absent.

To determine in which region the deformation was earlier, direct data must be sought outside the Gros Ventre–Hoback area. The Cache Creek anticline of the foreland area probably has its continuation in the Taylor Mountain anticline of the Teton Pass area.[19] Horberg[20] has found in that area that the Jackson thrust of the geosyncline partially covers the south limb of the anticline and that the foot-wall block of the Jackson fault west of Teton Pass is made up of beds ranging in age from Nugget to Upper Cretaceous, indicating considerable erosion of the Taylor Mountain fold

[19] Horberg, *op. cit.*, p. 33.

[20] *Ibid.*, pp. 35–36.

prior to thrusting. From this evidence it seems reasonably certain that the foreland structures developed somewhat earlier than the geosynclinal structures. This order of events seems quite logical also, inasmuch as the structures of the foreland area probably developed along earlier lines of weakness in the pre-Cambrian complex.

With the foreland structures developing first, they would serve as buttresses against which the geosynclinal structures might be pushed. Evidence of this buttressing effect is seen in the abrupt change of strike assumed by the outermost of the geosynclinal structures where they impinge on the south ends of the Gros Ventre and Teton ranges. The major structures of the Hoback Range trend essentially north-south until they approach the south end of the Gros Ventres, where they suddenly swing to the northwest around the buttress. In the same way the structures of the Snake River Range change from a north-south strike to a northwest strike where they meet the Teton structures. Further evidence of this buttressing effect is found in the numerous discontinuous thrusts developed at the northeast end of the Hobacks, where slicing occurred around the buttress.

In summarizing the Laramide diastrophism, the following series of events is indicated. Deformation may have started in the geosynclinal tract with downward movement, but such movement would not have to produce surficial structures to be seen in the field at the present. The earliest recognized deformation occurred in the foreland region. This movement developed the blocklike pattern so characteristic of the foreland region. It was first manifested at the surface by the formation of the asymmetrical anticlines. With continued stress these anticlines fractured and passed into well-developed trap-door blocks of ramp character. Within the Gros Ventre Range the maximum deformation occurred at the western end, as shown by the series of trap-door blocks becoming progressively stronger to the west; and it seems likely that the Cache Creek thrust fault is simply a trap-door structure in advanced stage of development, exposing the thrust-fault portion of the fracture.

The forces responsible for these structures were manifested largely in vertical movements, although some horizontal yielding is indicated. Somewhat later the deformation shifted to the geosynclinal area. Here the forces were most effective in the sediments themselves, producing shearing over the basement complex and crumpling the overlying strata into a series of close, tight folds. When the rocks could no longer yield sufficiently by folding, thrusting became dominant, forming the numerous thrust masses. It should be noted, however, that the sequence of events as here outlined applies to the border zone between the geosynclinal and foreland areas, making it quite possible for the earliest deformation in the central part of the geosynclinal tract to have started much earlier and even been simultaneous with the yielding in the foreland area. But, by the time the deformation had spread out to the margins of the geosyncline, the foreland area had become relatively stable. This would explain the lapse of time indicated between geosynclinal and foreland deformation within the area studied for this report. After this two-phase period of deformation, the foreland portion of the area (the Gros Ventres) apparently stood high, for detrital material (the Almy conglomerate) derived from it was deposited on the bordering structures of the geosynclinal tract.

Following the deposition of the Almy conglomerate the Laramide deformation came to a close with local renewal of thrusting on the Little Granite thrust, together with renewed movement on the previously outlined trap-door blocks.

TERTIARY FAULTING

A period of deformation, following at least the major phases of the Laramide orogeny, is recorded in the numerous high-angle faults which cut across the thrust sheets of the geosynclinal region. From evidence available within the area it is impossible to date these faults more definitely than to say that they are post-thrusting and, most of them, at least, post-Almy. The most important of these faults is the Hoback fault, which outlines the western border of the Hoback Range.

Relationships in the Teton Range show that during early Tertiary time Laramide structures were eroded to a surface of low relief, which was locally covered with volcanic materials, and that in early Miocene time a series of high-angle faults developed.[21] It is faulting of this later period which outlined and tilted the main Teton fault block. It seems logical that faulting of such magnitude near at hand should also be recorded in the mountainous area on the opposite side of Jackson Hole. Consequently, it is believed that the high-angle faulting of the Hoback area, which has a similar trend, developed at the same time and had its origin in the same forces as those responsible for the Tertiary faulting on the west side of Jackson Hole.

[21] F. M. Fryxell and Leland Horberg, personal communication.

EPEIROGENIC MOVEMENTS

The latest stages in the diastrophic history of the area are represented by widespread epeirogenic movements which caused stream rejuvenation and led to the carving of the present mountainous topography. These movements probably began late in the Pliocene and have continued intermittently to the present.

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THE RARER METALLIC CONSTITUENTS OF SOME AMERICAN IGNEOUS ROCKS. II[26]

ERNEST B. SANDELL
University of Minnesota

SAMUEL S. GOLDICH
Agricultural and Mechanical College of Texas

GEOCHEMICAL ASPECTS

GENERAL STATEMENT

The classical geochemical investigations of V. M. Goldschmidt[27] have thrown much light on the distribution and abundance of the chemical elements. Goldschmidt has shown that the distribution of many minor elements in igneous rocks is governed by the major constituents as a consequence of the isomorphous replacement of the latter in the crystal lattice of minerals when the ionic sizes of the two elements are sufficiently similar. Thus, nickel with an ionic radius of 0.78 A readily replaces magnesium, also having a radius o 0.78 A, in silicates. However, not only the ionic sizes but also the valences of the elements concerned are important in determining whether the minor element will enter the lattice of the major element mineral. Although lithium has the same ionic radius (0.78 A) as magnesium, it enters the early-crystallizing magnesium minerals to a very small extent, because the force acting on a univalent lithium ion is less than that acting on a divalent magnesium ion competing to enter the lattice. On the other hand, scandium, with an ionic radius of 0.83 A, shows a great tendency to enter ferromagnesian minerals, because, being trivalent, it is held with greater force than is magnesium or ferrous iron (ionic radius, 0.83 A). If the ionic sizes of two elements differ considerably, one will not replace the other. For example, barium (1.43 A) will replace potassium (1.33 A) but not sodium (0.98 A) to any appreciable extent.

In the discussion that follows, some possible relationships between the heavy metals and the major chemical constituents are considered. The ionic radii that are given are the empirical ones of Goldschmidt. The correlations that can be made between minor and major constituents lend much support to Goldschmidt's views, although a number of questions arise to which complete answers cannot be given. Probably the most significant of these is concerned with the presence of metals in forms other than isomorphous replacements of analogous elements in the lattices of the rock-forming minerals, as, for example, in sulphides. The entry of the minor elements, such as the heavy metals, into the crystal structure of the major-element minerals does not necessarily mean that these elements are fixed and entirely beyond the influence of later processes. Bowen has made an interesting case for the manifold possibilities of

[26] Continued from p. 115.

[27] For a partial list of the papers by V. M. Goldschmidt and his co-workers see the bibliography in "The Principles of Distribution of Chemical Elements in Minerals and Rocks," *Jour. Chem. Soc.*, 1937, p. 672. See, further, the series "Geochemische Verteilungsgesetze der Elemente," Vols. I–IX, *Skrifter Norske Videnskaps-Akad. Oslo I. Mat. Natur. Klasse, 1923–1937.*

magmatic differentiation by fractional crystallization. Reactions between early minerals and late liquids may change not only the composition of the solid phase but also that of the liquid phase by returning to the liquid elements no longer stable in the solid phase. These may be removed or may separate in new and entirely different forms. Likewise, even if the metals replace other elements, they will or may escape from magmas—for instance, in ore-forming emanations. These emanations as products of the magma will be largely determined by the course of the differentiation and the degree of fractionation that the liquid has undergone at different stages.

DISCUSSION OF ELEMENTS

MOLYBDENUM

Molybdenum has seldom been determined in igneous rocks. W. F. Hillebrand[28] was able to show the presence of this metal in a number of siliceous rocks, but he did not determine the amount present. J. B. Ferguson[29] found 0.01 per cent MoO_3 in a basalt from Hawaii. G. von Hevesy and R. Hobbie[30] reported the abundance of molybdenum in a composite of 282 eruptive rocks from central Europe as 0.0015 per cent and as 0.0003 per cent in a composite of 67 gabbros and norites. In the present investigation 22 determinations were made, and these average 0.00024 per cent molybdenum. The highest value (0.0007 per cent) was obtained from the granite-porphyry dike rock from Llano County, Texas. The average of 7 granitic rocks from central Texas is 0.0003 per cent. Granite from near Mine Lamotte, Missouri, contained 0.0002 per cent, and 6 granites from Minnesota averaged 0.00019 per cent molybdenum. The average percentage in 13 silicic rocks is 0.00025, and in 7 subsilicic rocks from Minnesota, 0.00021 per cent. An analysis of magnetite-ilmenite sand derived from the basic rocks of Minnesota showed 0.0013 per cent molybdenum. Although the average content of this metal is only slightly less in basic rocks than in acidic ones, the affinity of molybdenum for the siliceous granitic rocks is shown in its variation with silica content (Fig. 11).

BERYLLIUM

Beryllium has rarely been determined in igneous rocks. Most of our knowledge of the distribution of this element is based on investigations by Goldschmidt and co-workers.[31] These studies show beryllium to be concentrated in granites and especially in nepheline syenites. An average of 0.0007 per cent beryllium (0.002 per cent BeO) was found in 14 granites, and 0.0011 per cent in 22 leucocratic nepheline syenites. In nepheline the percentage of beryllium ranged from about 0.004 to 0.04 per cent (0.01–0.1 per cent BeO). Similar amounts were found in muscovite, albite, and tourmaline from granite pegmatites. Goldschmidt, Hauptmann, and Peters estimated the abundance of beryllium in the crust of the earth to be 0.0006 per cent on the

[28] "Distribution and Quantitative Occurrence of Vanadium and Molybdenum in Rocks of the United States," *Amer. Jour. Sci.*, Vol. VI (4th ser., 1898), p. 209; also *U.S. Geol. Bull. Surv. Bull. 167* (1900), p. 49.

[29] "The Occurrence of Molybdenum in Rocks with Special Reference to Those of Hawaii," *Amer. Jour. Sci.*, Vol. XXXVII (4th ser., 1914), p. 399.

[30] "Die Ermittlung des Molybdän- und Wolframgehaltes von Gesteinen," *Zeitschr. f. anorg. allgem. Chem.*, Vol. CCXII (1933), p. 134.

[31] V. M. Goldschmidt and C. Peters, "Zur Geochemie des Berylliums," *Nachr. Ges. Wiss. Göttingen Math.-physik. Klasse* (1932), p. 360; V. M. Goldschmidt, H. Hauptmann, and C. Peters, "Über die Berücksichtigung seltener Elemente bei Gesteinsanalysen," *Naturwiss.*, Vol. XXI (1933), p. 363.

basis of spectrographic determinations in sedimentary rocks.

The present investigation has been confined to a few silicic rocks; the analytical method used did not permit the accurate determination of the even smaller amounts of beryllium present in the subsilicic rocks. The results are given in Table 10. The average of 9 determinations, 1 representing a composite of 5 granitic rocks from Minnesota, is 0.0007 per cent. Like molybdenum, beryllium is concentrated in the silica- and alkali-rich differentiates of the magma; and a curve showing the variation of a beryllium with silica would show a sharp upward trend at a SiO_2 content of about 70 per cent. In the central Texas granites there is an unmistakable tendency for the two metals to maintain a constant ratio which is approximately Be/Mo = 2.5. The average beryllium content of these rocks is 0.0008 per cent; molybdenum, 0.00034 per cent. Although data for beryllium in basic igneous rocks are not available, it seems likely that the amount of beryllium in igneous rocks exceeds that of molybdenum.

TABLE 10

BERYLLIUM CONTENT OF ACIDIC IGNEOUS ROCKS*

DESCRIPTION AND LOCALITY	PER CENT	
	Be	SiO_2
Granite porphyry (Ilanite), Texas No. 5	0.0014	75.20
Granite, Bear Mountain, Texas No. 6	.001	76.77
Granite, Granite Mountain, Texas No. 3	.0008	73.02
Granite, Cassaday, Texas No. 4	.0005	72.15
Granite (quartz monzonite), Town Mountain, Texas No. 1	.0003	68.15
Granite, Graniteville, Missouri No. 2	.001	76.81
Rhyolite, Cook County, Minnesota No. 1	.0008	73.6
Granite, Southern Aitkin County, Minnesota No. 3	.0002	71.19
Composite of granitic rocks, Minnesota Nos. 2, 3, 4, 5, 6	0.0003	.....
Av. percentage	0.0007	.....

* E. B. Sandell, analyst.

LEAD

Studies of the lead content of igneous rocks have been made by Hillebrand,[32] J. D. Robertson,[33] J. B. Harrison,[34] A. M. Finlayson,[35] F. W. Clarke and G. Steiger,[36] J. A. Smythe,[37] and Hevesy and Hobbie.[38] Of these the work of Hevesy and Hobbie is the most comprehensive and reliable. Goldschmidt[39] states that potassium minerals, including the feldspars, contain from 0.0005 to 0.01 per cent lead. He has also shown that lead is not restricted in its occurrence to potassium minerals but may also occur in calcium minerals, as, for example, in apatite up to 0.005 per cent and in monoclinic pyroxenes in amounts from 0.0003 to 0.001 per cent. Goldschmidt expressed the opinion that a large proportion of lead is present as galena in igneous rocks.

Because of the similarity of ionic radii of lead and potassium (Pb, 1.32 A; K, 1.33 A) a relationship between these elements might be expected and is suggested by the data plotted in Figure 6. The preference of lead for silicic rocks may be largely explained by the enrichment of these rocks in potassium by processes of differentiation. Thus the

[32] "Geology and Mining Industry of Leadville," *U.S. Geol. Surv. Mono. XII* (1886), p. 591.

[33] *Mo. Geol. Surv.*, Vol. VII (1894), p. 480.

[34] *Report on Petrography of Cuyuni and Mazaruni Districts* (Georgetown: Demerara, 1905).

[35] "Problems of Ore-Deposition in the Lead and Zinc Veins of Great Britain," *Quart. Jour. Geol. Soc. London*, Vol. LXVI (1910), p. 299.

[36] "The Relative Abundance of Several Metallic Elements," *Jour. Wash. Acad. Sci.*, Vol. IV (1914), p. 58.

[37] *Trans. Nat. Hist. Soc. Northumberland, Durham and Newcastle-upon-Tyne* (1930).

[38] "Lead Content of Rocks," *Nature*, Vol. CXXVIII (1931), p. 1038.

[39] "Geochemische Verteilungsgesetze der Elemente, IX. Die Mengenverhältnisse der Elemente und der Atom-Arten," *op. cit.*, No. 4 (1937) p. 93.

lead content (0.0005 per cent) of the high-silica dike rock cutting the Skrainka diabase (Missouri samples Nos. 6 and 7) is influenced by the low K_2O content. The anomalous behavior of the felsites from Bare Hill and from Mount Houghton, Michigan, probably has some explanation in the magmatic history of these masses, but this explanation is not apparent in the chemical composition. The Medford diabase and the New Hampshire granite, although following the trend indicated by the bulk of the data, are somewhat deficient in lead with respect to both K_2O and SiO_2. No explanation of this behavior or of the fact that the granites from Bear Mountain, Texas, and from Graniteville, Missouri, are unusually rich in lead can be offered at this time. It should be noted, however, that these granites are remarkably similar in chemical composition.

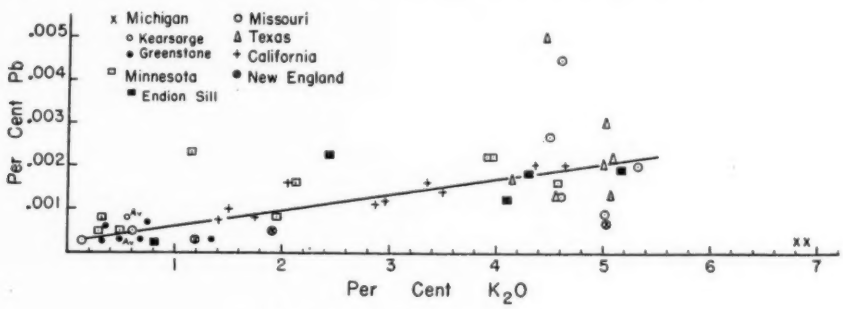

FIG. 6.—Variation of Pb with K_2O

The results of Hevesy and Hobbie and those of this study are in good agreement not only as to magnitude but also with respect to distribution. These investigators found 0.0005 per cent lead in a composite of 67 samples of gabbros and related types, but an average of 0.0010 per cent lead in a composite of 40 samples of essexites and related rocks. A composite of 26 samples of soda granites and soda syenites gave 0.0011 per cent lead, whereas a composite of 24 samples of potash granites and potash syenites averaged 0.0014 per cent. A composite of 58 samples of granitic rocks was found to contain 0.0030 per cent lead. In addition, an amphibolite inclusion in kimberlite from South Africa was found to contain 0.0015 per cent lead; kimberlite, from South Africa, 0.0016 per cent; lherzolite, from Maryland, 0.0019 per cent; and dunite, from North Carolina, 0.0042 per cent. The high lead content of these ultra-basic rocks is in sharp contrast to their low contents of potash and silica, and the distribution of the lead in these rocks deserves further study.

The average lead content of 220 igneous rocks is given as 0.0016 per cent by Hevesy and Hobbie. In the present investigation the average of 52 rocks is 0.0014 per cent. Twenty-nine silicic rocks ($SiO_2 > 63$ per cent) contained an average of 0.0019 per cent lead in comparison with 0.0009 per cent found for 22 basic rocks ($SiO_2 < 63$ per cent). Two dike rocks were not included in the above calculations. Five coarse-grained granites from central Texas gave an average of 0.0026 per cent lead, and 5 silicic rocks from Missouri, 0.0023 per cent.

ZINC

Reliable determinations of zinc in silicate rocks are few. Some values were obtained by Robertson[40] and by Finlayson,[41] and there are a number of others scattered through the literature. Clarke and Steiger[42] found 0.0064 per cent ZnO or 0.005 per cent zinc in a composite of 329 American igneous rocks. F. W. Clarke and H. S. Washington[43] gave 0.004 as the percentage of zinc in the igneous rocks of the crust of the earth. More recent data have been obtained by the Noddacks, who in 1930 estimated the abundance of zinc as 0.02 per cent[44] and in 1934 revised this figure to 0.007 per cent.[45] Goldschmidt[46] in unpublished investigations with N. H. Brundin and with H. Hörmann arrived at a percentage of not more than 0.004 for the abundance of zinc in the crust of the earth.

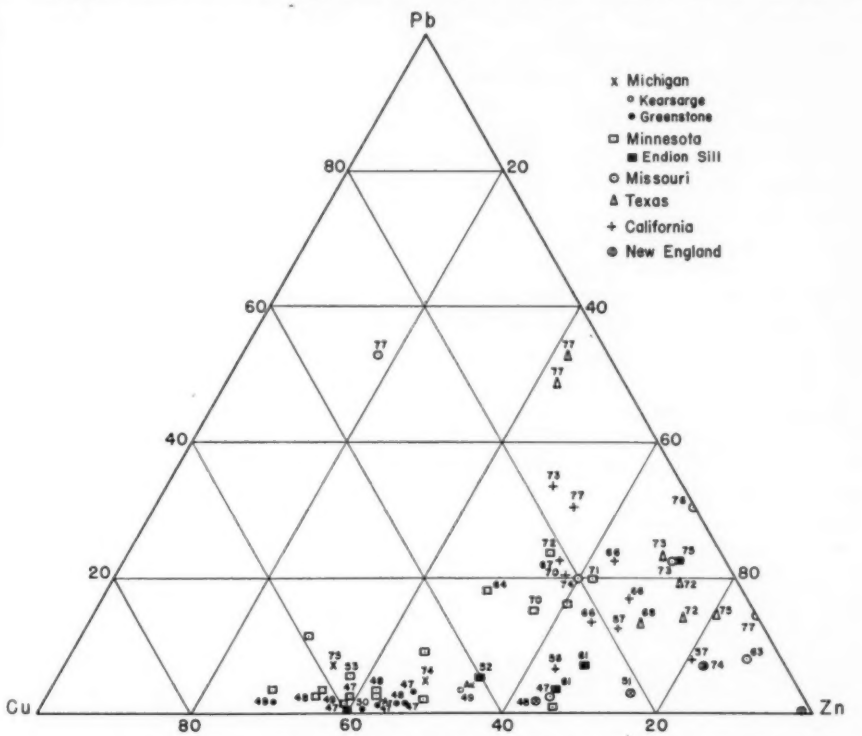

FIG. 7.—Relative proportions of Cu, Pb, and Zn in 58 igneous rocks. Approximate SiO_2 percentages of analyzed samples are given.

Although zinc occurs in largest amounts in basic rocks, it is commonly found in nearly as large quantities in the

[40] *Op. cit.*, p. 479.

[41] *Op. cit.*, p. 299.

[42] *Op. cit.*, p. 58.

[43] "The Composition of the Earth's Crust," *U.S. Geol. Surv. Prof. Paper 127* (1924), p. 20.

[44] I. and W. Noddack, "Die Häufigkeit der chemischen Elemente," *Naturwiss.*, Vol. XVIII (1930), p. 757.

[45] "Die geochemischen Verteilungskoeffizienten der Elemente," *Svensk Kem. Tid.*, Vol. XLVI (1934), p. 173.

[46] *Op. cit.*, p. 81.

silicic rocks. In the present study an average for zinc of 0.013 per cent was found for 25 basic rocks and of 0.0063 per cent for 29 acidic ones. In Figure 7 the relative proportions of copper, zinc, and lead for 58 rocks have been plotted together with the SiO_2 percentages for most of the samples. The predominance of zinc can be seen in the grouping of the points in the diagram. In general it can be said that zinc occupies a position intermediate between copper and lead in its distribution. Thus, in Figure 7 the analyzed samples may be divided into three major groups: (1) a copper-zinc group; (2) a zinc group with almost equal proportions of copper and lead, but with the latter two metals clearly subordinate to the zinc; and (3) a zinc-lead group. In the first of these groups lead usually constitutes less than 10 per cent by weight of the three metals and averages about 5 per cent. These samples are chiefly subsilicic rocks ranging from 47 to about 60 per cent SiO_2. In the lower portion of this range copper dominates and in a large number of the samples composes 50 per cent or more of the total weight of the metals. With increasing silica content zinc becomes more important. The second group of samples is dominated by zinc, ranging from 50 to 90 per cent of the three metals. These samples range roughly from 60 to 75 per cent in SiO_2. The third group of samples characterized by low copper percentages includes the highly siliceous igneous rocks generally with SiO_2 contents of about 73 per cent or more. In several of these samples lead is the dominant metal.

Goldschmidt[47] states that the investigations of Brundin show that zinc enters the silicates of iron and magnesium, such as augite, hornblende, and especially biotite. This is well corroborated by tests made on biotite from the Texas granites and on hornblende from the New Hampshire granite. In addition, zinc can be expected in magnetite and in ilmenite. A magnetite-ilmenite concentrate derived from the basic rocks of Minnesota was found to contain 0.03–0.04 per cent zinc. That the distribution of zinc should be influenced by the distribution of iron is understandable from the very similar ionic radii of zinc and divalent iron (0.83 A). In correlating zinc with ferrous iron, however, it was plainly evident that in a number of the samples oxidation of the ferrous iron had taken place. For this reason zinc has been plotted against total iron calculated as Fe_2O_3 (Fig. 8). The zinc content increases rapidly with iron in the range from 0 to 4 per cent Fe_2O_3. Considerable scattering of the points in this range is introduced by the Texas and Missouri granites. In the range from 4 to 18 per cent Fe_2O_3 the majority of the samples fall fairly close to the curve; but the Endion sill samples, the rhyolite from Iron Mountain, Missouri, and the basic phase of the Mount Bohemia gabbro are marked exceptions. These samples, together with the granites mentioned, seem to require a second curve, which has been shown as a broken line in Figure 8. It is obvious that other factors besides the total iron content influence the zinc content of the samples.

The ratio MnO/Fe_2O_3 is linear, or nearly so, over a wide range of iron. Accordingly, a plot of zinc against MnO approaches linearity with a ratio of Zn/MnO of approximately 0.07/1 for rocks with more than 0.1 per cent MnO. In the more silicic rocks with less than 0.1 per cent MnO, the ratio of zinc to MnO is considerably higher. Because the ionic radius of manganese (Mn^{II}, 0.91 A) ap-

[47] *Op. cit.*, p. 6.

proaches that of ferrous iron, it replaces this element in silicates; and, like iron, manganese is a rough index of zinc in igneous rocks. It appears that iron and manganese are removed in proportionately slightly larger amounts than zinc in the early crystals during differentiation, and zinc is thus enriched with respect to content, although the aplite (No. 7) does not conform. The same tendency is noticeable in other samples.

Although zinc and cobalt have similar ionic radii and the same valences, they do not behave alike. Cobalt is most abundant in magnesia-rich rocks, but zinc will not favor such rocks unless iron

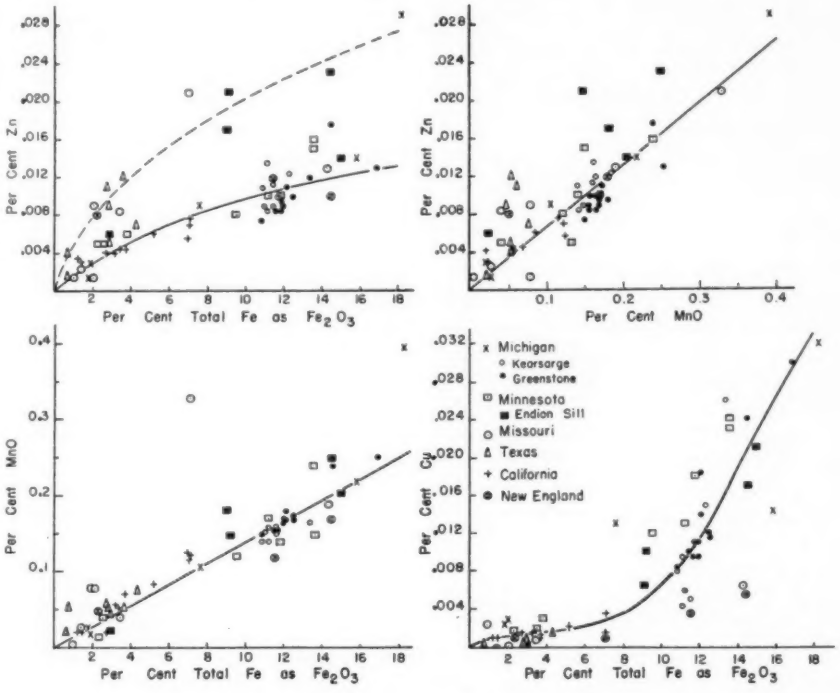

FIG. 8.—Variations of Zn, MnO, and Cu with total iron (Fe_2O_3) and of Zn with MnO

these two metals in the residual liquid. When the latter finally crystallizes, zinc accompanies iron (and manganese) into biotite. It should follow that the more silicic the rock, the larger, in general, should be the ratio Zn/Fe. There is considerable evidence in the case of the Texas granites that there is an upward trend in the ratio with increasing silica is abundant at the same time. In the granitic rocks, which normally are low in magnesia, cobalt is rare, but zinc is commonly present in relatively large quantities.

CADMIUM

The results of determinations of cadmium in 9 samples, 3 of which are composites, are given in Table 11. The

amounts of this element in igneous rocks are so minute that the analytical values have a high percentage of error, and therefore no attempt has been made to trace the relationships of cadmium to the major chemical constituents. The close affinities of cadmium for zinc have long been known, and Clarke and Washington[48] based their estimate ($n \times 10^{-7}$ gm. of cadmium per gram of rock) of the abundance of this element in the crust of the earth on the work of Clarke and Steiger[49] showing that in zinc ores the ratio of zinc to cadmium is approximatly 200/1. This ratio, according to J. H. L. Vogt,[50] ranges from 600/1 to 1,000/1. More recently Goldschmidt[51] has set the value of the ratio Zn/Cd = 90/1. In the present study a fairly constant ratio of cadmium to zinc was found, as the values given in Table 11 show. Although these ratios are given to two significant figures, the last figure is not highly accurate. The average ratio as Cd/Zn is 0.002/1, or the ratio of Zn/Cd is 500/1.

TABLE 11

CADMIUM CONTENT OF IGNEOUS ROCKS*

DESCRIPTION AND LOCALITY	PER CENT Cd	Cd/Zn
Granite, Granite Mountain, Texas No. 3	0.00002	0.0022
Granite, southern Aitkin County, Minnesota No. 3	.000015	.003
Granodiorite, St. Cloud, Minnesota No. 7	.00001	.0017
Graphic granite, Kanabec County, Minnesota	none	.....
Gabbro, Duluth, Minnesota No. 12	.00001	.0010
Diabase, Cook County, Minnesota No. 16	.000025	.0021
Composite of 5 granites, central Texas	.00001	.0016
Composite of 5 granites, Minnesota	.000013	.0022
Composite of 5 basic rocks, Minnesota	0.00002	0.0017

* E. B. Sandell, analyst.

The data indicate that cadmium is more abundant in basic than in acidic rocks. The size of the cadmium ion is such (1.03 A) that replacement of calcium (1.06 A) by cadmium might be expected; and, as a matter of fact, Goldschmidt found about 0.0005 per cent cadmium in some Norwegian labradorite rocks. It also seems possible that cadmium, like zinc, is concentrated in ferromagnesian minerals, the amount depending upon the iron content. In granites and in related rocks the major part of the cadmium no doubt occurs in biotite. There is some indication that the ratio Cd/Zn is slightly larger in acidic than in basic rocks, as would be expected, since the ionic size of cadmium is greater than that of zinc and proportionately more of the former element should therefore remain in the mother-liquid.

COPPER

Clarke and Steiger[52] found 0.0117 per cent of CuO or 0.009 per cent copper in a composite of 329 American igneous rocks, and Steiger[53] found 0.0155 per cent copper in a composite of 71 Hawaiian lavas. According to Harrison,[54] the average copper content of 36 igneous and metamorphic rocks of British Guiana is 0.025 per cent. R. C. Wells[55] found 0.034 per cent copper in a sample of Columbia River basalt. F. F. Grout[56] analyzed some Keweenawan diabases from Minnesota and found copper to range from 0.029 to 0.032 per cent. Broderick[57] re-

[48] *Op. cit.*, p. 21.

[49] *Op. cit.*, p. 61.

[50] "Ueber die Concentration des ursprünglich fein vertheilten Metallgehaltes zu Erzlagerstätten," *Zeitschr. f. prakt. Geol.*, 1898, p. 386.

[51] *Op. cit.*, p. 83.

[52] *Op. cit.*, p. 58.

[53] "Data of Geochemistry," *U.S. Geol. Surv. Bull. 770* (1924), p. 641.

[54] *Op. cit.*

[55] "Data of Geochemistry," *U.S. Geol. Surv. Bull. 770* (1924), p. 641, n. 5.

[56] "Keweenawan Copper Deposits," *Econ. Geol.*, Vol. V (1910), p. 471.

[57] "Differentiation of Lavas . . . . ," *op. cit.*, p. 555; "Differentiation in Traps," *op. cit.*, p. 303.

ported 0.009 per cent and 0.012 per cent as the weighted averages of copper for the Kearsarge and the Greenstone flows. These values were duplicated in the present study. The average copper content of the Keweenawan traps of the Michigan copper district is given as 0.01 per cent by Broderick, and a range of from 0.002 to 0.038 per cent copper was found for the igneous rocks of this region. Other determinations of copper in igneous rocks have yielded values of similar magnitude. Clarke and Washington[58] estimated the copper content of the crust of the earth as 0.010 per cent, and this figure has been accepted by Goldschmidt.[59]

The well-known tendency for copper to occur in largest amounts in basic rocks is illustrated in the present study. Twenty silicic rocks (average SiO_2, 72 per cent) had an average copper content of 0.0016 per cent; 11 intermediate rocks (average SiO_2, 62 per cent) contained 0.0038 per cent; and 15 subsilicic rocks (average SiO_2, 48.5 per cent) contained 0.0149 per cent. The basic igneous rocks contain about ten times more copper than the acidic ones. There are good reasons for accepting the observation that copper usually occurs in igneous rocks largely as sulphides or as alteration products of original sulphide minerals. This is particularly true of the basic rocks. The relationship between copper and sulphur in the flows of Michigan has been discussed in some detail, and it seems probable that copper appropriates sulphur from the magma to meet its needs so far as a supply is available. However, a high sulphur content is not a criterion of a high copper content. The diabase from Skrainka, Missouri, may be cited to illustrate this point. This rock, with a relatively high sulphur content of 0.10 per cent, contains 0.0065 per cent of copper. Similarly, low copper and high sulphur characterize the samples of the Medford diabase.

Grout[60] has suggested that some of the copper in igneous rocks may be present in silicate minerals, and the same suggestion is made by J. C. Reed[61] on the basis of comparative chemical and petrographic studies of a basic rock from Alaska. The chemical analyses show about five times as much copper as does the microscope. Goldschmidt has assigned identical ionic radii to Cu^{II} and Fe^{II} (0.83 A), so that it seems plausible that copper enter the ferromagnesian silicates. The variation of copper with total iron (Fe_2O_3) is shown in Figure 8.

NICKEL

The nickel content of igneous rocks and the distribution of this metal in rock-forming minerals have been discussed by Vogt.[62] He concluded that, although some nickel may be present in sulphides in common igneous rocks, most of the nickel is contained in the silicate minerals, particularly the ferromagnesian minerals. The feldspars and the feldspathoids carry little or no nickel. The most important of the nickel-bearing silicates is olivine, some analyzed specimens containing as much as 0.5 per cent NiO. The nickel content increases with the content of Mg_2SiO_4 in olivines. The average of 13 analyses of iron-poor olivines (MgO, 47–51 per cent) is given by Vogt as 0.33 per cent NiO. Four iron-rich olivines contained an average of 0.21 per

[58] *Op. cit.*, p. 20.

[59] *Op. cit.*, p. 72.

[60] *Op. cit.*, p. 475.

[61] "Nickel Content of an Alaskan Basic Rock," *U.S. Geol. Surv. Bull. 897-D* (1939).

[62] "Nickel in Igneous Rocks," *Econ. Geol.*, Vol. XVIII (1923), pp. 307–53.

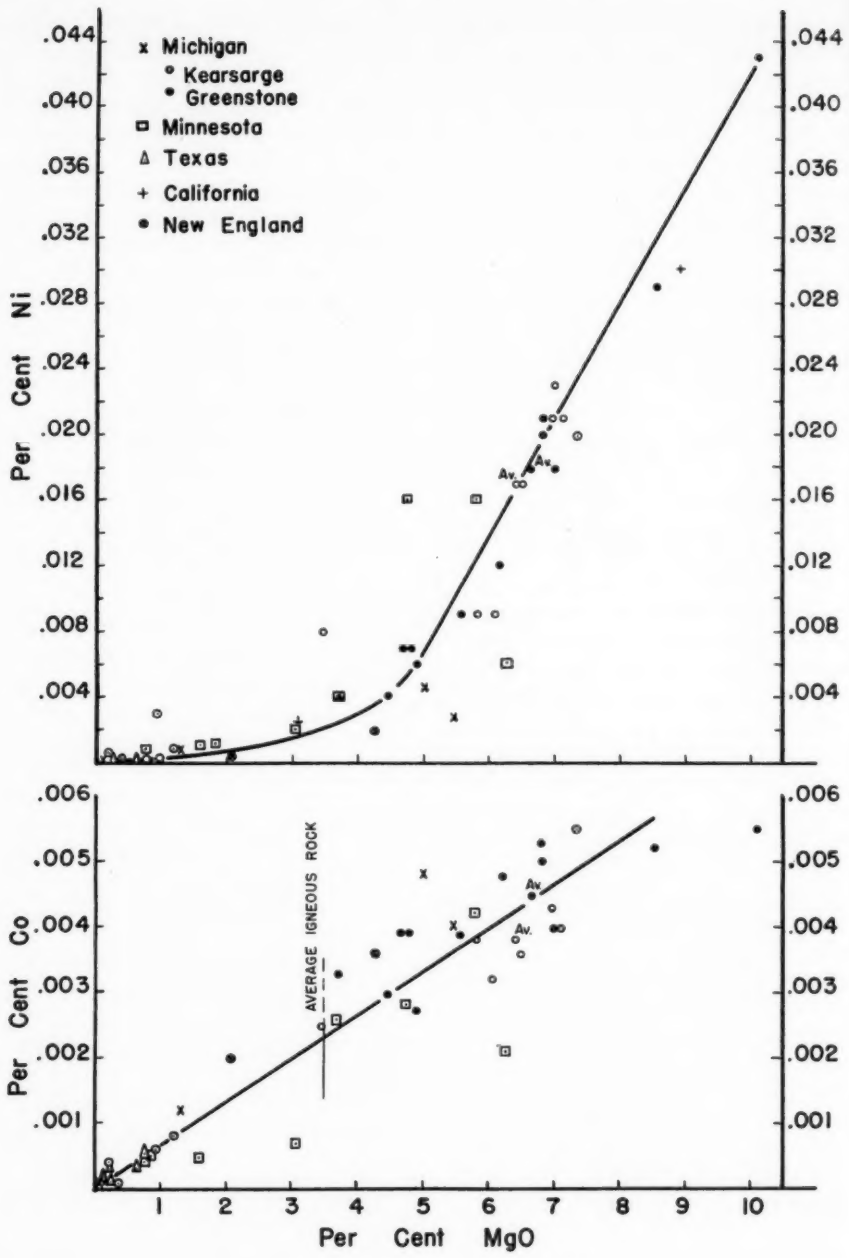

FIG. 9.—Variation of Ni and Co with MgO

cent NiO. Two new analyses are recorded in Table 12.

TABLE 12

CHEMICAL ANALYSES OF OLIVINES*

	1	2
SiO_2	40.90	41.11
TiO_2	0.05	0.02
Al_2O_3	0.22	0.00
Cr_2O_3	0.02	tr.
Fe_2O_3	0.59	0.49
FeO	8.24	6.38
MnO	0.12	0.11
NiO	0.30	0.39
MgO	49.78	51.43
CaO	0.05	0.07
H_2O+	0.00	0.00
H_2O-	0.01	0.02
Total	100.28	100.02
Sp. gr. $t°/4°$	3.321	3.309

* S. S. Goldich, analyst. Analyses made in the Department of Geology, Washington University, St. Louis, 1934.
1. Olivine, Rice Station School, Talkai, Arizona. (U.S. Nat. Mus. No. 86128.) Fo 86.5, Fa 11.7, Δ1.8. (Weight per cent.)
2. Olivine, Kragerö, Norway. Fo 89.3, Fa 9.1, Δ1.6.

Following olivine in order of decreasing nickel contents are the orthorhombic pyroxenes, the monoclinic pyroxenes, amphiboles, and biotite. Nickel in granitic rocks is probably largely restricted to biotite. The order given above is that of Bowen's[63] reaction series, and a close relationship between the distribution of nickel and crystallization differentiation is indicated, as has been pointed out by Vogt. The following figures for nickel in rock groups are given by Vogt: peridotites, 0.14 per cent, pyroxenites, 0.08 per cent; all facies of norites, 0.035 per cent; olivine-free gabbros, diabases, basalts, and related rocks, 0.030 per cent; syenites, monzonites, diorites, and related rocks, 0.006 per cent; and granites, less than 0.001 per cent. Additional values for groups of rocks have been published recently by Goldschmidt, and these are included in Table 13. Vogt computed an average of 0.0093 per cent of nickel for plutonic igneous rocks. His estimate of 0.01 per cent for the abundance of nickel in the igneous rocks of the crust of the earth is undoubtedly more reliable than the figure of 0.020 per cent given by Clarke and Washington.[64] The work of Goldschmidt, Witte, and Hörmann[65] also sets the percentage of nickel in the crust of the earth at 0.01 per cent.

The almost identical ionic radii of magnesium and nickel (0.78 A) explain the close association of the two elements. In Figure 9, nickel has been plotted against the MgO contents of the samples investigated in the present study. The nickel content rises rapidly in the range above 4 per cent MgO, indicating the tendency for nickel to be enriched in the early fractions of the magnesian or ferromagnesian minerals. This can be expressed in the relation

$$\frac{[\text{Ni}]\ \text{solid}}{[\text{MgO}]\ \text{solid}} = k\,\frac{[\text{Ni}]\ \text{liquid}}{[\text{MgO}]\ \text{liquid}}, \quad (1)$$

in which the distribution coefficient k must have a value greater than 1. As crystallization proceeds, the residual liquid becomes impoverished in nickel more rapidly than in magnesium, and the last crystals to separate will have a lower ratio of Ni/MgO than the first. On the assumption that equation (1) can be applied and that k remains more or less constant during crystallization, an attempt may be made to obtain an approximate value for the coefficient from the relation between nickel and magnesia in the samples of the Greenstone flow of Michigan. The weighted average percentages for nickel and for MgO in the flow are 0.018 and 6.65, and these values are assumed to approximate the concen-

[63] *The Evolution of the Igneous Rocks* (Princeton: Princeton University Press, 1928), p. 60.

[64] *Op. cit.*, p. 20.

[65] *Op. cit.*, p. 67.

trations of these constituents in the original magma. Accordingly, the curve of Figure 10 plotted for the Greenstone flow samples may be considered to consist of two portions, one representing the Ni/MgO ratio of samples containing less magnesia than the average for the flow, and the other containing more magnesia. Then the lower portion of the curve extending from the point for average Ni/MgO for the flow to the origin of the graph represents samples which have had their magnesia contents reduced by sinking of early magnesia-rich crystals, notably olivine. The upper portion of the curve represents samples enriched by the addition of the olivine crystals. This concept is similar to the idea suggested in Vogt's[66] *proto-enriched* and *eutect-enriched* magmas.

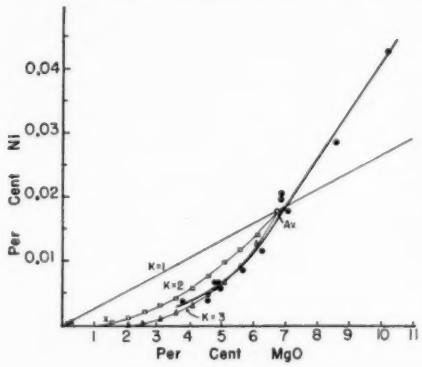

FIG. 10.—Variation of Ni with MgO in the Greenstone flow, Michigan, and theoretical curves showing variation of Ni with MgO in crystallizing liquid.

Three theoretical curves are also shown in Figure 10. These give the magnesia and nickel contents of the residual liquid after removal of various amounts of magnesia by crystal settling. The curve $k = 1$ shows the composition of liquid if the ratio of nickel in the solid and in the liquid phase remained the same, and under these conditions both early and late crystals would show the same percentage of nickel. The curves $k = 2$ and $k = 3$ are based upon calculations in which the distributive coefficient k in equation (1) is assumed to be 2 in the one case and 3 in the other. It is seen that the Greenstone curve follows approximately the theoretical curve for $k = 3$. There are no data to indicate the trend of the Greenstone curve near the origin, but the Ni-MgO curve (Fig. 9), based on additional samples, indicates that an extension of the Greenstone curve in Figure 10 would probably pass above the theoretical curves $k = 2$ and $k = 3$. Perhaps this is to be explained by a decrease in the value of k at lower MgO concentrations. It is probable that the value of the coefficient depends upon the nature of the mineral crystallizing, and the difference might be large, for example, between olivine and biotite.

No well-defined empirical Ni-MgO curve can be expected. Considerable scattering of points may be anticipated, especially in the upper portion of the curve. Thus a peridotite and a pyroxenite having about the same magnesia content should show different nickel contents, the former the higher one. The principle may have considerable importance in the genesis of nickel deposits, and further reference will be made to it in a later section.

COBALT

A relation between cobalt (0.83 A) and magnesium (0.78 A) is more apparent than one between cobalt and ferrous iron (0.83 A). This relation between cobalt and magnesium, in contrast to that between nickel and magnesium, is linear over a wide magnesia range (Fig. 9). This does not mean that there is no re-

[66] "Nickel in Igneous Rocks," *op. cit.*, p. 326.

placement of ferrous iron by cobalt. (For example, this is known to take place in magnetite-ilmenite.)

Because nickel is strongly enriched in the early crystal fractions, the ratio Ni/Co is much greater in subsilicic than in silicic rocks. In the latter, cobalt may even predominate over nickel. Early estimates of the ratio Ni/Co in igneous rocks are too high, because they were based on determinations of these elements in basic igneous rocks. The ratio Ni/Co obtained in this study is approximately 3.5/1. Goldschmidt has assigned the value 2.5 to this ratio. It is of interest to compare the average values for the cobalt contents of various rock groups as given by Goldschmidt with the predicted content calculated on the assumption that the ratio Co/MgO is constant and equal to about 0.00066, the average value found in this study. The predicted values and the observed values are given in Table 13.

TABLE 13

OBSERVED AND PREDICTED COBALT VALUES FOR IGNEOUS ROCKS

Rock	Per Cent			
	MgO*	Ni*	Co*	Calculated Co
Peridotite	40	0.31	0.024	0.026
Gabbro	8	.016	.008	.005
Diorite	3	.004	.003	.002
Granite	1	.0002	.0008	.0007
Nepheline syenite	1	0.0002	0.0008	0.0007

* Adapted from figures given by Goldschmidt, *Jour. Chem. Soc., London*, 1937, p. 662.

ABUNDANCE OF THE MINOR ELEMENTS IN THE CRUST OF THE EARTH

Although the primary purpose of this investigation was not to determine the average content of the metals considered in the crust of the earth, the data obtained allow some averages to be calculated. Values for the abundance of the elements have generally been obtained by averaging the individual percentages of many samples assumed to represent the known rocks or by determining the percentages of the elements in a composite of many samples. The first method was used by Clarke, Washington, and others; the second, by Goldschmidt, Hevesy, and by others as well. The comparatively small number of samples analyzed in this study makes these methods somewhat uncertain, and an attempt has been made to arrive at values for the minor metals by using two indirect methods:

1. If the mass relation between the minor element and a major constituent is linear or approximately so, the value of the ratio and the known average percentage of the major element in the crust of the earth permit the calculation of the average percentage of the minor element.

2. If the empirical relation between the percentages of the minor element and of silica can be established, the average percentage of the former can be calculated from the relative abundance of rocks in the lithosphere in the known silica range. For this purpose, the silica range has been broken up arbitrarily into units of 5 per cent. The relative abundance of rocks in each unit was obtained by computations based chiefly upon the data given by W. A. Richardson and G. Sneesby[67] for the frequency distribution of silica in igneous rocks.

Method 1 is preferable to method 2 but is less generally applicable. The satisfactory application of method 2 re-

[67] "The Frequency-Distribution of Igneous Rocks. I. Frequency-Distribution of the Major Oxides in Analyses of Igneous Rocks," *Min. Mag.*, Vol. XIX (1922), p. 306.

quires that a fairly constant ratio of metal to silica exists, at least over the critical portion of the range. This condition is not always well fulfilled, and the reliability of the silica-metal curve may be open to question if it is based on relatively few points. Nevertheless, a better average probably can be computed in this manner than by simply taking the arithmetical average. The application of the two methods may be illustrated by the calculation of the abundance of cobalt:

Method 1.—Cobalt varies linearly with magnesia, and the value of Co/MgO is 0.00066 (Fig. 9). The percentage of magnesia in the lithosphere is 3.5, and, accordingly, the percentage of cobalt is 0.0023.

Method 2.—The computations are shown in Table 14.

TABLE 14

AVERAGE PERCENTAGE OF COBALT IN THE EARTH'S CRUST FROM ABUNDANCE IN ROCKS OF VARIOUS SILICA CONTENTS

SiO_2 Percentage Range	Estimated Per Cent of Lithosphere	Percentage of Cobalt Contributed
<40.0...	0.5	0.005×0.01 =0.00005
40.0–47.5....	13.0	0.13 ×0.005 =0.00065
47.5–52.5....	17.0	0.17 ×0.0035=0.00060
52.5–57.5....	17.0	0.17 ×0.0025=0.00043
57.5–62.5....	12.0	0.12 ×0.002 =0.00024
62.5–67.5....	12.0	0.12 ×0.001 =0.00012
67.5–72.5....	13.0	0.13 ×0.0007=0.00009
72.5–80.0....	13.0	0.13 ×0.0002=0.00003
>80.0....	2.5	0.025× nil = nil
Av. percentage of cobalt, 0.0022		

The average percentage of nickel calculated from the nickel-silica relationship is 0.008. This average is more uncertain than the one for cobalt, because of the relatively large amount of nickel in the most basic rocks whose abundance and nickel content are difficult to estimate.

Estimates of the abundance of lead may be made from the lead-silica relation and from the lead-potash relation. Calculations based on these curves give an identical value of 0.0014 per cent. However, the Pb-K_2O curve does not pass through the origin, indicating that not all the lead present in igneous rocks replaces potassium; and, accordingly, the average of 0.0014 per cent for lead should be increased somewhat. The uncertainty in the slope of the Pb-K_2O line is so great that the application of a correction hardly seems justified, but the value for lead may be rounded off to 0.0015 per cent.

If it is assumed that the relation between zinc and MnO is represented by a straight line passing through the origin, the average zinc content of igneous rocks is 0.008 per cent, based on a MnO content of 0.124 per cent given by Clarke and Washington. This figure for zinc, however, is probably a little low, because the silicic rocks contain more zinc than can be accounted for by their MnO contents.

The calculation of the abundance of copper on the principles mentioned above is difficult because there is a wide range in the copper content of rocks of about the same silica content. A value of 0.007 per cent was obtained as a weighted average for copper for the samples analyzed (Table 16). This figure may or may not approximate the average percentage of copper in the crust of the earth, depending largely upon how closely the copper content of the igneous rocks of Minnesota and of Michigan approach the world average.

The abundance of the minor metals as found in the present study, together with the results of what are regarded as the most reliable previous investigations, is

summarized in Table 15. There is substantial agreement in the values for lead, copper, and nickel. The new value for zinc is twice as large as the older one. The most serious difference is in the figures for molybdenum; the new value is only one-sixth as great as the earlier value.

TABLE 15

ABUNDANCE OF MINOR ELEMENTS IN IGNEOUS ROCKS

Metal	Per Cent	
	Previous Value	Present Study
Mo	0.0015*	0.00025
Be	.0006†	< .0007
Pb	.0016‡	.0015
Zn	.004§	.008
Cd	.00005‖	.000015
Cu	.01¶	.007
Ni	.01**	.008
Co	0.004**	0.0023

* Hevesy and Hobbie, *op. cit.*, p. 134.
† Goldschmidt, Hauptmann, and Peters, *op. cit.*, p. 363.
‡ Hevesy, *Fortschr. Min. Krist., u. Petrog.*, Vol. XVI (1932), p. 147.
§ Unpublished data of V. M. Goldschmidt, N. H. Brundin, and H. Hörmann mentioned by Goldschmidt, *op. cit.*, p. 81.
‖ Unpublished results of Goldschmidt and Hörmann mentioned by Goldschmidt, *ibid.*, p. 82.
¶ Clarke and Washington, *op. cit.*, p. 20.
** Data of Goldschmidt, Witte, and Hörmann, mentioned by Goldschmidt, *op. cit.*, p. 67.

GEOLOGIC IMPLICATIONS

GENERAL STATEMENT

In this section it is proposed to review the data for their possible regional significance and for their geologic implications.

KINDS OF IGNEOUS ROCK AND MINERALIZATION

The concept of a genetic relation of ore deposition to magmatic processes stimulated efforts to associate certain kinds of mineralization with specific rock types or with groups of rocks. Grout's[68] "tree" is an attempt to show the association of ore deposits with igneous rock clans. Numerous other investigators have given attention to this problem, and the subject is admirably treated in a paper by A. F. Buddington[69] in which the many complexities of the problem are discussed. Paul Niggli[70] has placed somewhat different emphasis on the association of ores with igneous rocks. He states that certain ore deposits, like igneous rocks, must be regarded as the products of magmatic differentiation, and in this sense he finds Spurr's term "ore magma" useful. Niggli writes:

. . . . Regarding, as we do, the individual igneous rock types as products of such a differentiation, we shall, in general, be concerned not with the strict co-ordination of *certain ore deposits and certain rocks, but rather with the question whether certain courses of orthomagmatic differentiation constantly give rise to certain types of ore deposits.* Two products of a differentiation process may not be singled out from the rest; the phenomenon must be taken as a whole.[71]

DISCUSSION OF DATA

In Table 16 averages for the metals calculated for the different regions are given. For the purpose of this tabular summary the rocks from each area have been divided into two groups, silicic and subsilicic, the division being made at 63 per cent SiO_2. Although the distribution of the rarer metallic constituents in igneous rocks cannot be explained solely on the basis of a silicity principle, the variations of the metals with silica have been plotted in Figures 11 and 12. Geologists are familiar with this type of diagram,

[68] *Petrography and Petrology* (New York: McGraw-Hill Book Co., 1932), p. 140.

[69] "Correlation of Kinds of Igneous Rocks with Kinds of Mineralization," *Ore Deposits of the Western States* (Amer. Inst. Min. Met. Eng., 1935), pp. 350–86.

[70] *Ore Deposits of Magmatic Origin* (London: Thomas Murby & Co., 1929), trans. H. C. Boydell.

[71] *Ibid.*, p. 69.

and the variation curves should be helpful in relating the distribution of the metallogenic elements to igneous-rock types and probably also in suggesting how this distribution may be influenced by processes of magmatic differentiation. The points for the detail curves showing nickel, cobalt, total iron (Fe_2O_3), zinc, larly for the Michigan and Minnesota series, are based on a much greater number of samples than have been analyzed for their heavy-metal content.

DISTRIBUTION OF METALS

Lead and molybdenum.—The behavior of lead and of molybdenum needs no

TABLE 16

SUMMARY OF AVERAGES FOR HEAVY METALS IN GROUPS OF ROCKS

	Per Cent						
	SiO_2	Mo	Pb	Zn	Cu	Ni	Co
A. Silicic rocks (SiO_2>63 per cent):							
Texas Nos. 1–6	73.19	0.00031	0.0025	0.0080	0.0010	0.00019	0.00026
Missouri Nos. 1–5	73.17	.0002	.0023	.0085	.0009	.00036	.00027
California Nos. 6–11, 13	69.16	.....	.0016	.0042	.0013	.....	.....
Minnesota Nos. 1–7, 22	72.27	0.00019	.0020	.0063	.0043	.00103	.00041
Michigan Nos. 3156, 3157	74.03	.....	.0003	.0023	.0028	.00045	0.00015
New England No. 3	74.36	.....	0.0007	0.0080	0.0010	.....	.....
Weighted av.	71.93	0.00025	0.0019	0.0063	0.0021	0.00058	0.00030
No. of samples	26	13	29	29	29	19	20
B. Subsilicic rocks (SiO_2<63 per cent):							
Missouri No. 7	47.27	.....	0.0005	0.0130	0.0065	0.020	0.0055
California Nos. 1–3	56.77	.....	.0008	.0067	.0020	.....	.....
Minnesota Nos. 8–21	51.73	0.00021	.0010	.0140	.0152	.0110	.0028
Michigan Nos. 3161, 3163, 3166, Kearsarge and Greenstone	50.23	.....	.0007	.0146	.0162	.0086	.0037
New England Nos. 1–2	49.52	0.00015	0.0004	0.0110	0.0045	0.0013	0.0028
Weighted av.	51.67	0.0002	0.0009	0.0130	0.0126	0.0097	0.0032
No. of samples	20	8	22	25	25	17	16
Av. $\frac{A+B}{2}$	61.80	0.00023	0.0014	0.0097	0.0074	0.0051	0.0018

and copper in the Kearsarge and the Greenstone flows have been purposely omitted, but the weighted averages are shown. Figure 13 is a composite of variation diagrams showing the Keweenawan igneous rocks of the Michigan copper district after Broderick, the Duluth sills and overlying flows after Schwartz and Sandberg, the Clear Lake area lavas after Anderson, and the central Texas granites. The variation curves, particu- special comment, except to point out the marked affinity of molybdenum for the high-silica granites indicated by the pronounced upward swing of the Mo-SiO_2 line at a silica content of about 72 per cent.

Nickel and cobalt.—These metals, closely related in their distribution, clearly favor the basic igneous rocks. From the diagram it can be seen that high nickel contents are restricted to

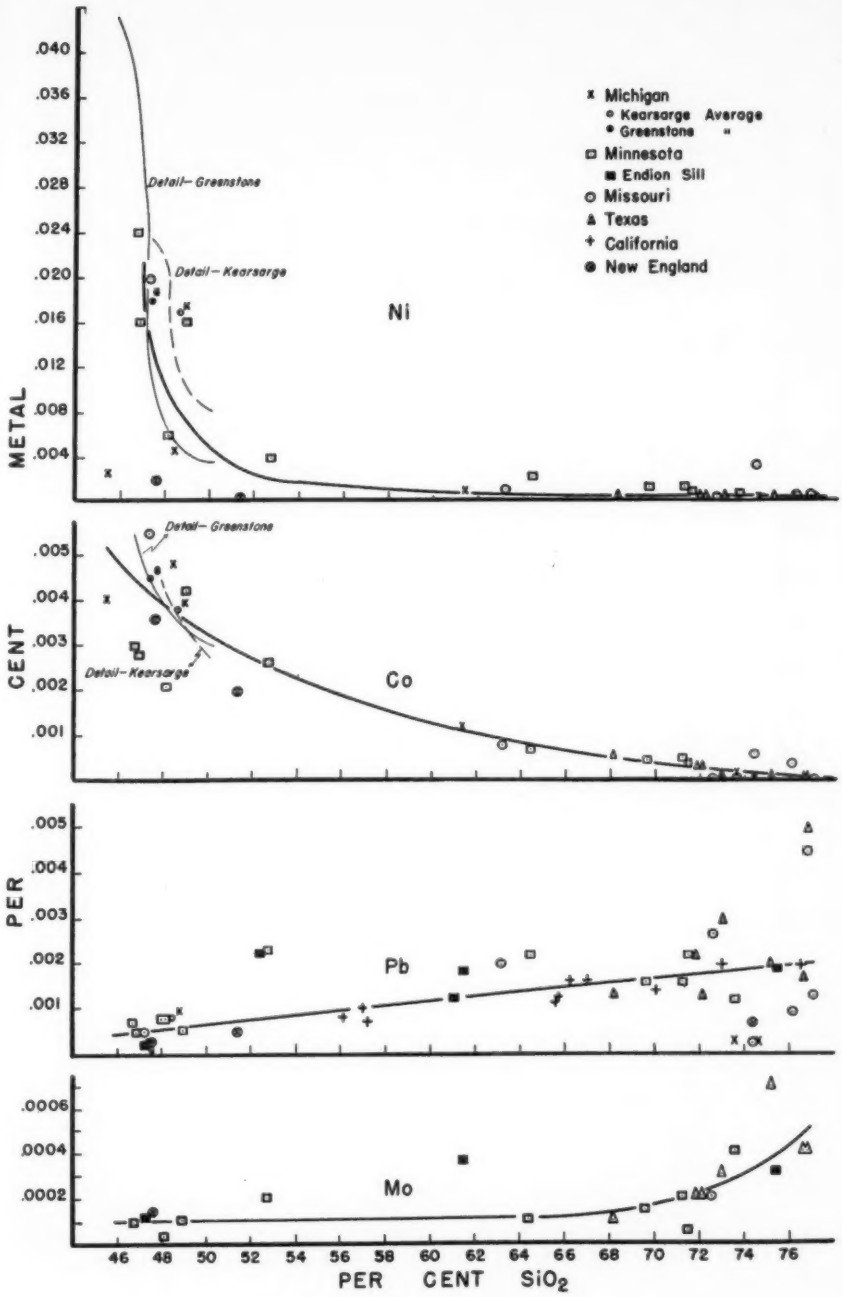

FIG. 11.—Variation of Ni, Co, Pb, and Mo with SiO_2

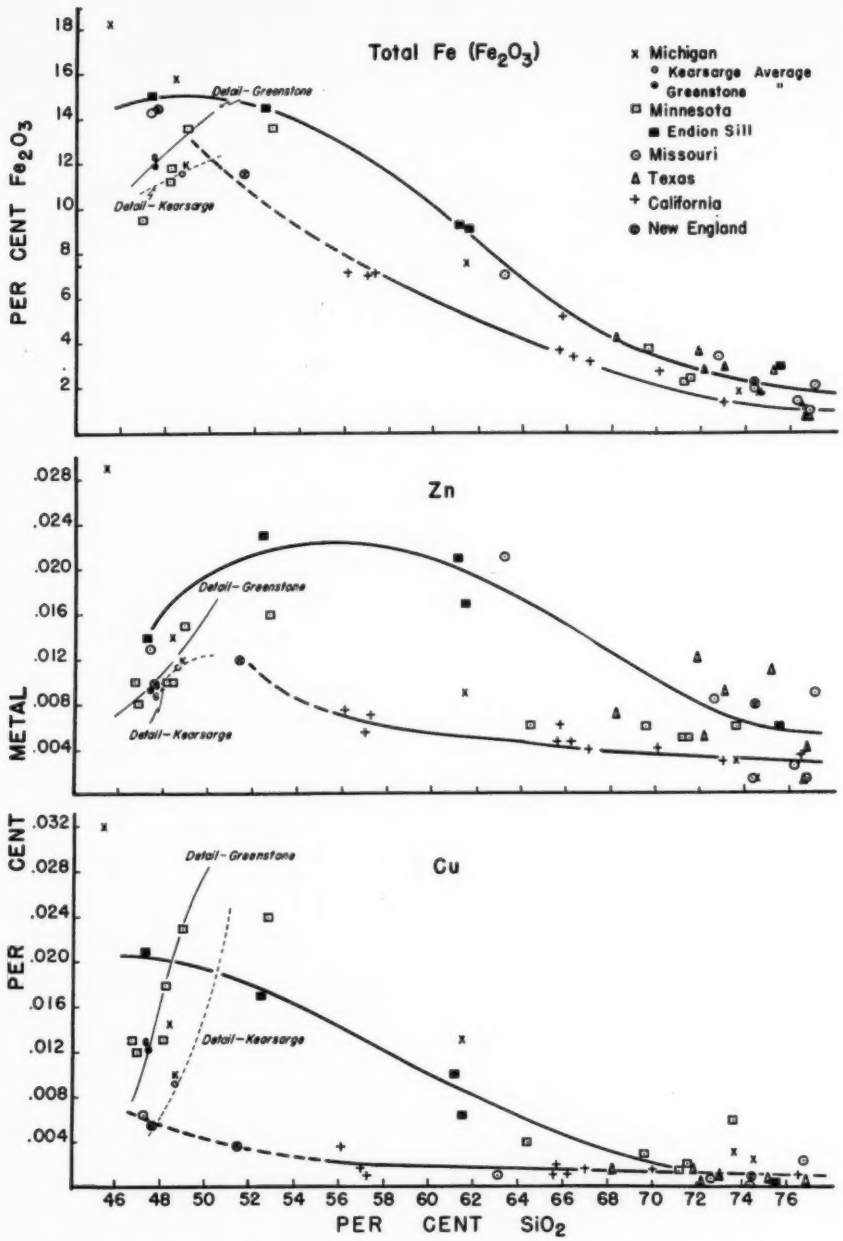

FIG. 12.—Variation of total iron (Fe_2O_3), Zn, and Cu with SiO_2

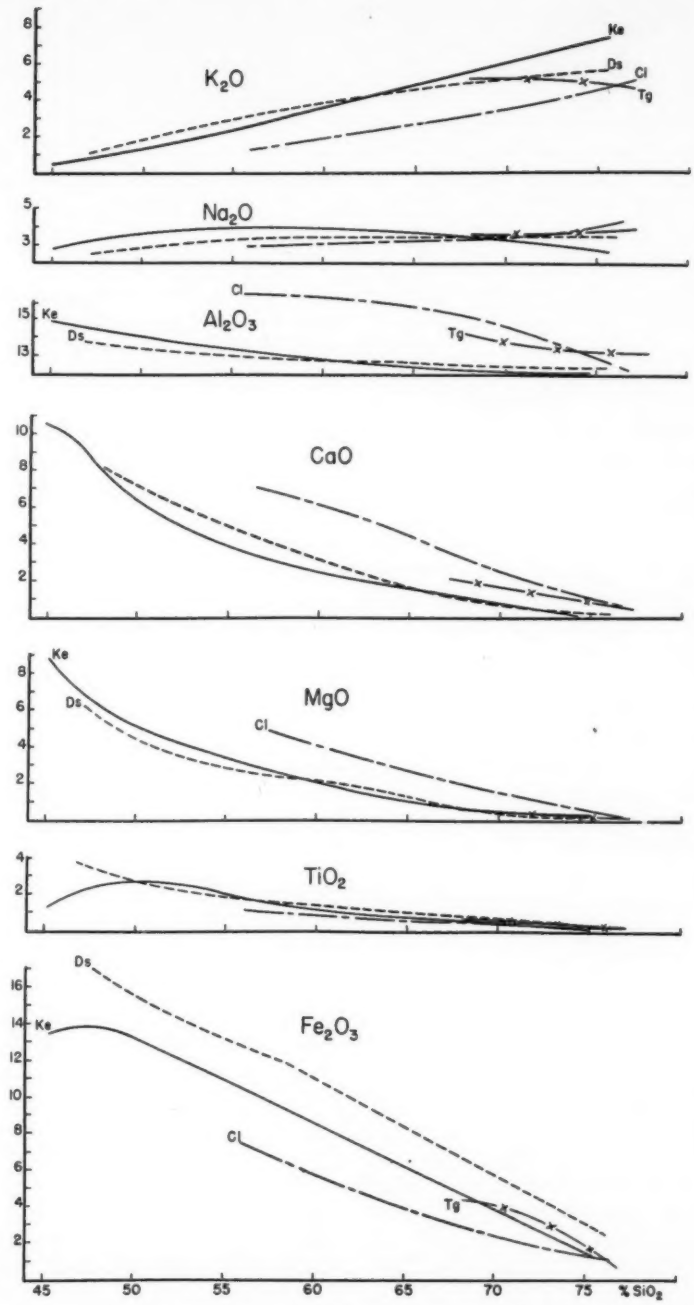

FIG. 13.—Comparison of variation diagrams. *Ke*, Michigan copper district (*solid line*); *Ds*, Duluth sills, Minnesota (*dash line*); *Cl*, Clear Lake Area, California (*dot-dash line*); *Tg*, central Texas (*x-dash line*). Per cent of oxides plotted on vertical scale against per cent of silica on horizontal scale.

rocks with less than 50 per cent SiO_2. Both these metals are rare in granitic rocks. The one exception represents the soda granite dike cutting, and probably genetically related to, the diabase at Skrainka, Missouri.

Copper and zinc.—The distribution of copper and of zinc affords some special problems, and detailed consideration should be given to certain features.

If the samples from Michigan and Minnesota can be said to be representative of the Lake Superior region, this area is characterized by the relatively high copper content of its igneous rocks. This is true of both acidic and basic varieties, and the percentage of copper is from three to four times as great as that of similar rocks from the other regions studied. Two curves show the variations of copper in Figure 12. The lower curve has been drawn for points representing the Clear Lake area of California. This curve and a tentative extension represent fairly well the Missouri and Texas samples, the Medford diabase, and the granite from New Hampshire. A second curve is needed to fit the higher copper values of the samples from the Endion sill, Minnesota, and this curve approaches Broderick's curve for the variation of copper with silica in the Michigan copper district.

Considered together, the variation curves Cu/Fe_2O_3 (Fig. 8) and Cu/SiO_2 are instructive, because, if copper is influenced by the iron content, the flows of the Clear Lake area have copper percentages in keeping with their tenor of iron, but the New England rocks and the diabase from Skrainka, Missouri, show deficiencies of copper. Thus, although regional differences of the nature of petrographic and metallogenic provinces are indicated, it is probable that some of these differences reflect different lines of magmatic descent or simply different magmatic histories.

Zinc variations for the Endion sill and the California volcanics, like those of copper, can be represented in two distinct curves. As in the case of copper, these variations are in part influenced by the behavior of iron. The Zn/Fe_2O_3 curves (Fig. 8) show that the Clear Lake area flows have zinc contents in keeping with their iron contents but that certain granites from Texas and Missouri, the granite from New Hampshire, the red rock facies of the Endion sill, and the basic phase of the Mount Bohemia gabbro have a greater ratio of zinc to Fe_2O_3. These samples fit the higher of the two curves drawn to represent the variations of total iron with silica in the analyzed samples (Fig. 12). In the higher silica range it appears that zinc is more nearly controlled by iron than is copper, but the reverse seems to be true of the lower silica range. Thus, a line connecting the weighted averages for zinc in the Greenstone and the Kearsarge flows slopes upward and is roughly parallel to the trend shown in the detail curves for zinc and also in the Endion sill in this silica range. A line connecting the two samples of the Medford diabase reveals the same trend; however, similar lines for copper and total iron all have downward slopes with increasing silica or trends which are quite different from the variations of copper and iron shown by the details for the Michigan flows.

MAGMATIC PROCESSES

The samples from the Endion sill and from the flows of the Michigan copper district represent facies enriched or impoverished in femic and calcic minerals by crystal settling and further changed through local and upward concentration of mineralizers, possibly also through

gaseous transfer. Through these processes the basaltic magmas of the Kearsarge and Greenstone flows suffered strong fractionation, which was carried even further in the Duluth sills producing the facies known as the "red rock." This same type of differentiation operated on a larger scale in the Duluth lopolith, yielding even more varied rock types; and undoubtedly analogous processes were effective in the Sudbury lopolith and in other masses. In contrast, the samples of the Clear Lake area of California represent separate flows, and the chemical analyses of these can be said to approach the compositions of a series of liquids. The relation of the flows to each other is not readily determinable, and there is no conclusive evidence that the flows are all differentiates of the same parent-magma. Thus, rock series such as are represented in the Endion sill and Michigan flow facies are not strictly comparable to a series of volcanics such as the Clear Lake lavas, and the observed differences probably reflect to a large extent the different magmatic histories of the samples involved rather than regional differences of the nature of petrographic provinces or epochs. The possibilities of such differences should be entertained, but more data and additional groups of rocks are needed to fully test them.

Considered as a whole, the data suggest that rock facies produced by differentiation must be examined in the light of their geologic occurrence as much as in terms of chemical differences, and a great variety of rock types and rock series should be expected as a result of modifications imposed upon crystallization differentiation by physical conditions. It is interesting to note that in published studies of three groups of samples—namely, the flows from Michigan, the Endion sill, and the granites from central Texas—special attention has been given to the probable effect of the mineralizers and their role in the differentiation, although some important differences in emphasis are apparent.

Broderick[72] has placed great emphasis upon gaseous transfer as the mechanism producing the differentiation in the Michigan flows. The distribution of cobalt and of nickel in the flows, however, shows that crystallization differentiation with crystal settling cannot be ignored. If the distribution of copper, zinc, and other constituents is to be explained by gaseous transfer, this mechanism obviously did not become effective before fractional crystallization of the magma had proceeded to such an extent that the nickel was largely tied up in early olivine crystals. Thus, without attempting to detract from the excellent study made by Broderick, the question must be raised as to just how much silicate transfer by volatiles, especially in the gas phase, has occurred and to what extent there has been "silicate transfer of the volatiles."[73]

Crystallization of early magnesium-rich olivine will rob a magma of its nickel and, to a lesser extent, of its cobalt. Vogt has called attention to the close relationship that must exist between crystallization differentiation resulting in norites and the genesis of nickel deposits. Although space does not permit discussion of Vogt's detailed theory, a few points deserve mention. Thus Vogt fully appreciated the fact that nickel would be concentrated in the early fraction of olivine crystals separating from the magma. Therefore he postulated that these crystals settled and were remelted to give a proto-enriched magma from which the

[72] *Op. cit.*, pp. 503–58.

[73] Bowen, *The Evolution of the Igneous Rocks*, p. 296.

nickel separated as a liquid sulphide phase or that the nickel was transferred from the silicate compound to the sulphide compound owing to a chemical reaction which he illustrated as follows: "FeS + Ni-cation in silicate compound $\rightleftharpoons$ NiS + Fe-cation in silicate-compound."[74] Because of the fact that nickel shows a lesser tendency to be concentrated in the orthorhombic pyroxenes than in olivine, reaction between olivine and liquid to produce bronzite or hypersthene may return nickel to the magma, thus making it available for further concentration into ore deposits. Further speculations along these lines are probably unwarranted. A. Knopf,[75] for instance, has concluded that the Sudbury nickel deposits are a replacement controlled by fractures and are considerably younger than the norite.

CONCLUSIONS

1. The distribution of copper, zinc, nickel, and cobalt in the Kearsarge and Greenstone flows of Michigan can be correlated with the distribution of the major chemical constituents. Nickel and cobalt show a close relationship to MgO and CaO, whereas copper and zinc appear to be related to the distribution of iron.

2. Features similar to those that characterize the Michigan flows are found in the Endion sill of Minnesota. The copper content of the igneous rocks from Minnesota, both silicic and subsilicic, is from three to four times as great as that of similar rocks from other regions, with the exception of the Michigan copper district.

3. The marked similarities in the bulk chemical analyses of the granitic rocks from the St. Francois Mountains, of Missouri, and from the Llano uplift, of Texas, are reflected in the heavy metal content of these rocks. Some averages for comparison are: SiO_2: Texas, 73.2 per cent; Missouri, 73.2 per cent; Minnesota, 72.3 per cent. Lead: Texas, 0.0025 per cent; Missouri, 0.0023 per cent; Minnesota, 0.0020 per cent. Zinc: Texas, 0.008 per cent; Missouri, 0.009 per cent; Minnesota, 0.006 per cent. Copper: Texas, 0.001 per cent; Missouri, 0.001 per cent; Minnesota, 0.004 per cent. Nickel: Texas, 0.0002 per cent; Missouri, 0.0004 per cent; Minnesota, 0.0010 per cent. Cobalt: Texas, 0.0003 per cent; Missouri, 0.0003 per cent; Minnesota, 0.0004 per cent.

4. Copper, lead, and zinc were determined in a series of 10 samples from the Clear Lake area, California. The results plotted in a number of diagrams gave surprisingly smooth curves.

5. A granophyric facies of the Medford diabase was found to contain more lead and zinc than the normal phase but less copper, nickel, and cobalt. Analyses of a granite from the White Mountain district of New Hampshire and of soda amphibole separated from this rock show that all the zinc present in the granite can be accounted for by the amphibole but only small fractions of the copper and lead.

6. Molybdenum and beryllium show a strong affinity for the high-silica granites. This is in accordance with the known tendency for these elements to be concentrated in the pegmatitic differentiates.

7. Lead in igneous rocks shows a close relationship to potash, although all the lead is not present in potassium minerals. Some may be present in galena, as has been indicated by Goldschmidt.

8. Zinc shows a strong affinity for iron and enters minerals such as magnetite-

[74] "Nickel in Igneous Rocks," *op. cit.*, p. 253.

[75] "Petrology," chapter in *Geology, 1888–1938, Fiftieth Anniversary Volume* (New York: Geol. Soc. Amer., 1941), p. 347.

ilmenite, pyroxenes, amphiboles, and biotites. Additional factors, little known, appear to influence its distribution. Manganous oxide is an index element of zinc in igneous rocks.

9. The average ratio of Zn/Cd was found in the present study to be 500/1.

10. Copper, like zinc, is influenced by the distribution of iron, although, in contrast to zinc which is commonly abundant in granitic rocks, copper shows a preference for the subsilicic ones. The presence of copper chiefly in the form of sulphide minerals, especially in the basic igneous rocks, is well established. Factors other than the sulphur content of the magma, however, probably determine the copper content. Thus, a high-sulphur content is not a criterion of a high-copper content in an igneous rock.

11. Nickel clearly favors the basic, magnesium-rich igneous rocks, and olivine is the chief nickel-bearing silicate. The early fractions of the ferromagnesian minerals tend to be enriched in nickel, and as a result the nickel content of olivines increases with the content of Mg_2SiO_4.

12. Cobalt varies linearly with magnesia, and the ratio of Co/MgO is approximately 0.00066/1. This value may be used to predict the cobalt content of common igneous rocks with a fair degree of accuracy. From this relation the average percentage of cobalt in the igneous rocks of the crust of the earth may be set at 0.0023 per cent.

13. A second method for estimating the abundance of a minor element in the crust of the earth is based upon the empirical relation between SiO_2 and the minor element and on the relative abundance of rocks in the lithosphere in portions of the known silica range. Values for the probable abundance of the heavy metals in the crust of the earth have been obtained by these methods.

14. Along the lines of the present investigation further studies are needed not only for groups of rocks but also for the component minerals of these rocks. Once information is available indicating the control exercised by processes of magmatic differentiation (these processes must also be studied) in the distribution of the metallic constituents of the magma, the groundwork may be set for a more intelligent attack on the equally difficult problem of how these metals are concentrated, transported, and deposited to form ore bodies.

ACKNOWLEDGMENTS.—Material assistance was received from E. D. Burr, L. Gullings, R. Nordgren, and Donald Smith, who, as students in the School of Chemistry, of the University of Minnesota, contributed a number of chemical analyses and heavy-metal determinations. A large number of the samples used in this study were supplied by T. M. Broderick, G. M. Schwartz, A. E. Sandberg, Carl Tolman, C. A. Anderson, and R. I. Dickey. The generous cooperation of these persons and of F. F. Grout and R. B. Ellestad, of the Rock Analysis Laboratory, is gratefully acknowledged.

FLOW STRUCTURES IN THE BEREA SANDSTONE AND BEDFORD SHALE OF CENTRAL OHIO

JOHN R. COOPER
Ohio Wesleyan University

ABSTRACT

The structures in the Berea sandstone and the Bedford shale described in the past as "concretionary" or "disturbed" layers are shown to result from contemporaneous mass movements on the free surface of deposition. The top of the disturbed rock was generally truncated by erosion before the overlying undisturbed beds were laid down. Lateral movements are evident in many instances and at many levels in the formations. These movements and others in which lateral movement is not readily apparent are interpreted as surface flows and slides. The causes of subaqueous flows are reviewed in an attempt to find the reason for the flows. The conclusion is reached that they may have started spontaneously in the soft sediment as it accumulated under the conditions which obtained. Important causes of such flows include an increasing weight of sediment, a slight surface slope, and occasional beds or seams of clay which flowed readily and also acted as a lubricant between sand layers, permitting them to slide past one another.

INTRODUCTION

This paper deals with certain irregularities in the Berea sandstone and Bedford shale of Delaware and Franklin counties in central Ohio. The irregular portions of the Berea sandstone, known to quarrymen of the past generation as "nigger head" or "turtle-back" stone, were described as "concretionary" by N. H. Winchell[1] and Edward Orton.[2] Of one occurrence Orton wrote: "The only explanation suggested is the vague one that the rock was wrought into these shapes by concretionary force."[3] Since that time they have generally been called "concretionary" or "disturbed" structures.

Bedding planes with ripple marks and cross-bedding are tilted, folded, and broken in the irregular portions—facts which prove that once horizontal beds have been disturbed in a mass movement. It is in these movements that the origin of the "concretionary" structure is to be found.

[1] "Report on the Geology of Delaware County," *Rept. Geol. Surv. Ohio*, Vol. II (1874), pp. 272–313.

[2] "Report on the Geology of Franklin County," *Rept. Geol. Surv. Ohio*, Vol. III (1878), pp. 596–616.

[3] *Ibid.*, p. 638.

Two main questions arise: (1) Did the movement occur on the surface of deposition before the overlying beds were laid down, or did it take place beneath more or less cover as a between-the-beds movement? (2) What cause or causes lie back of the movements? Before either of these questions is considered, it will be necessary to review the stratigraphic section and describe several localities.

LOCALITIES STUDIED

The localities studied are along the Berea sandstone and Bedford shale outcrop northeast of Columbus (Fig. 1). A survey of the literature indicates that similar structures are common in these formations from the Ohio River to the northern part of the state.

STRATIGRAPHY

The two formations involved lie at the base of the Waverly series of lower Mississippian age. From below upward, they are Bedford shale and Berea sandstone.

BEDFORD SHALE

The Bedford, approximately 90 feet thick, is a laminated clay shale, for the most part gray but in part red. Thin,

ripple-marked layers of fine-grained sandstone are interbedded with the shale in the top 20–25 feet of the formation. They are more abundant and thicker near the top, but even at the very top shale predominates, the sandstone being confined to scattered beds 1–2 inches thick. Fossils prove that the Bedford is a marine deposit.

BEREA SANDSTONE

The Berea is a medium- to fine-grained sandstone about 40 feet thick. Typically it is medium- to thick-bedded in the middle and upper part, with shaly layers toward the base. Clay shale is present as partings and beds up to 1 foot thick. It is more abundant toward the base. Many of the sandstone beds are lenticular, and gently dipping cross-bedding and ripple marks are common. The ripple marks are mostly of the symmetrical, wave-formed type and show that the deposit was laid down under water, presumably under the sea. Fossils are very rare in the Berea, none having been found in the area studied.

An unconformity is recognized between the two formations at many places in northern Ohio where the Berea lies on a highly irregular surface of the Bedford.[4] In the area studied as the basis of this paper, no such unconformity is recognizable. The two formations pass into each other by alternation through a vertical distance of a few feet, and it is impossible to put the finger on a plane of separation. The same seems to be true for most of central and southern Ohio. To be sure, C. R. Stauffer has described an unconformity at Lithopolis, 15 miles southeast of Columbus.[5] The Berea, here only 4 feet thick, rests on an irregular surface of the Bedford. The unconformity indicated may be only a local phenomenon like the numerous cases of scour and fill found within the Berea itself.

ROCKY FORK LOCALITY

The best-exposed section in the area studied is found along Rocky Fork, east of Gahanna, 9 miles northeast of Columbus. The section exposed at locality *C* of Figure 1 is indicated in Table 1.

TABLE 1

SECTION EXPOSED IN THE "LONG CLIFF" AT ROCKY FORK

		Feet
Sunbury...	Black fissile shale, exposed	2
Berea.....	Massive sandstone, beds not well defined	8½
	Lenticular-bedded sandstone, beds, averaging 6 inches thick	11
	Fine shaly sandstone	2½
	Sandstone flow layer	4
	Even-bedded sandstone in layers 2 inches–1 foot thick, with shaly beds and some clay shale	8
	Clay shale with thin, ripple-marked sandstone	3
	Persistent sandstone ("first sand") 8 inches thick, twisted and broken at four places	1
Bedford...	Clay shale with thin, ripple-marked sandstone, exposed	3

The first good sandstone layer met in going up from the Bedford is taken as the basal member of the Berea and is referred to as the "first sand." Four localities in the ravine are of special interest.

TREE-CLIFF SECTION (LOCALITY A)

This is on the north side of the brook where the Berea sandstone is first seen

[4] W. G. Burroughs, "The Unconformity between the Bedford and Berea Formations of Northern Ohio," *Jour. Geol.*, Vol. XIX (1911), pp. 655–59; H. P. Cushing, Frank Leverett, and F. R. Van Horn, "Geology and Mineral Resources of the Cleveland District, Ohio," *U.S. Geol. Surv. Bull. 818* (1931), pp. 47–48.

[5] *U.S. Geol. Surv. Folio 197, Columbus Folio* (1915), p. 11 and Pl. V.

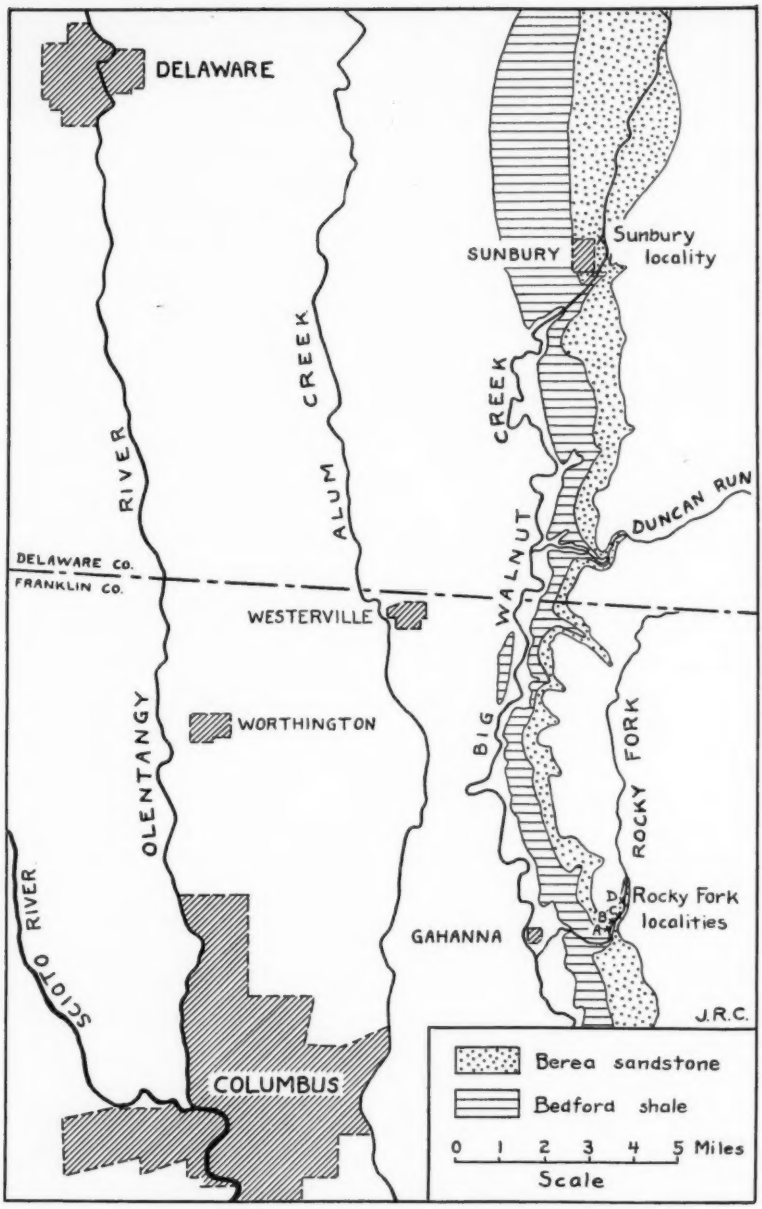

FIG. 1.—Index map of localities studied

upstream from Gahanna. The lower 25 feet of the cliff is Bedford, the upper part Berea. Disturbance by mass movement is well shown, involving the Bedford and the lower part of the Berea (Fig. 2). The disturbed rocks appear in a flaring U-shaped area, 120 feet across and at least 25 feet deep. The shale is crumpled and faulted, and the first sand is broken into segments which are folded and thrown into gnarled balls. The contorted rock, including segments of the first sand, is truncated sharply by a nearly horizontal surface and overlain by 5 feet of hard shale, 4 feet of interbedded shale and sandstone, and then the medium-bedded sandstone of the typical Berea. The lower boundary of the disturbed mass is well defined, being characterized at some places by a clay seam and fault displacement. The movement was evidently a slump accompanied by flowage of the shale. Because of the details of the upper contact, it is thought that the slump took place on the sea bottom and all surface irregularities were eroded off before the hard shale was deposited.

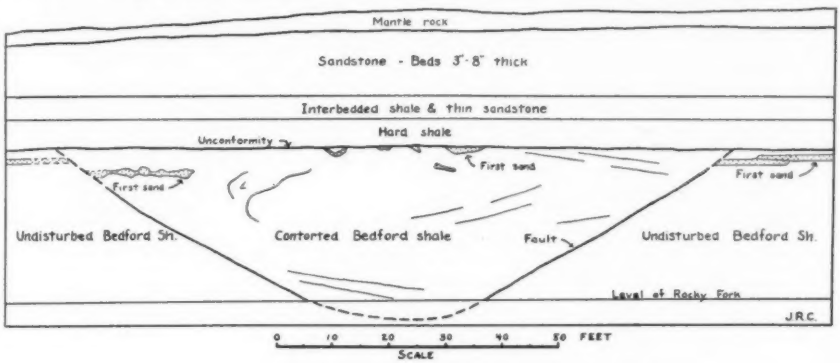

FIG. 2.—Field sketch of the tree-cliff section, Rocky Fork

NORTH-BANK SECTION (LOCALITY B)

The first sand can be traced upvalley until it runs below stream-level. About a quarter of a mile up from the tree cliff, it is exposed in the north bank, where it is more or less broken and twisted and at one point folded back on itself (Figs. 3 and 4).

The exposed length of this recumbent fold is 15 feet. The shale below it is little disturbed. At several places less than 1 foot below the first sand, thin ripple-marked sandstone layers prove that the shale is in normal position, the top of the beds facing upward. A 1-inch rippled sandstone is found likewise, wrapping around the blunt end of the fold. The ripple marks on this bed, which is only 2 inches from the first sand, show that the top of the bed faces the first sand. Thus it is clear that the fold is a syncline. The shale above the fold is much disturbed and contains gnarled fragments and balls of sandstone. No ripple marks were found on any of these sandstone fragments.

An interesting feature of the fold is that the tip end is somewhat thinned and turned back on itself for a distance of $4\frac{1}{2}$ feet. The stone in this turned-back portion is gnarled or kneaded in appearance.

The explanation suggested is that, soon after the first sand was deposited and while it was still plastic, it was

caught up in laterally flowing mud and rolled back on itself with a rolling hinge-motion. From the fact that the fold sags downward it is probable that some of the mud flowed out from beneath the sand. The continued growth of this kind of fold seems to require the presence of mud between the limbs to act as a lubricant. Where this is lacking, as it is near the axis of the fold under consideration, the limbs may be rolled back together.

turbed stretches range in length from 14 to 60 feet. The associated shale is involved also, the disturbed mass commonly appearing as a broad V, much wider at the top than near the bottom. The top is truncated by flat-lying, undisturbed shale, 2 inches–2 feet above the first sand; the bottom is in every instance below stream-level and therefore not exposed.

The most probable explanation is that

FIG. 3.—North-bank section at Rocky Fork, showing the first sand folded back on itself

LONG-CLIFF SECTION (LOCALITY C)

A short distance upstream from the last locality and in the south bank there is a cliff over 500 feet long. At the west end the first sand is exposed a few feet above stream-level. It is separated from the overlying sandstone beds by 3 or 4 feet of shale containing thin, ripple-marked sandstone layers, the latter generally more abundant toward the top. For most of the length of the cliff these layers and the first sand, 8 inches thick, run along undisturbed. But at four places the first sand is buckled and broken into contorted fragments. The disa considerable area of the soft sediment, only a few feet thick, started to flow laterally on the sea floor, the first sand being carried forward on a very plastic base of mud. The first sand, lacking the fluidity of the associated mud, was locally buckled. It seems likely that the underlying muds pierced these buckled places and, welling up from below, spread out on the sea floor.

EAST-END SECTION (LOCALITY D)

From the long cliff northward to the highway bridge about half a mile upstream many disturbed layers in the

body of the Berea sandstone are exposed. They are not so spectacular as those just described but are of a more common, widespread type. In fact, disturbances like them—the so-called "concretionary structures"—are found in the Berea sandstone wherever it is exposed in central Ohio.

The disturbances are confined to layers a few inches to a few feet thick but of considerable areal extent. Commonly a 20-foot cliff will show several of them separated by undisturbed beds. Sometimes a single, disturbed layer may be traced 100 feet or more before it dies out. In the meanwhile another may have appeared above or below it.

Where individual beds are discernible in the disturbed layers, they are more or less folded. Small open folds, closed folds, and overturned folds have been observed, as well as what appear to be portions of fan folds and one example where a layer of sandstone was rolled up like a jelly roll or a snowball rolled over wet snow. In some cases the beds are twisted into a confused mass, in which no system is apparent. The bottom of the disturbed layers is gradational, whereas the top is sharply truncated and overlain by undisturbed beds.

Photographs taken in the east end of Rocky Fork will make the relations clearer. Figure 5 shows a typical flow layer seen in cross section, and Figure 6 the top of another flow layer exhumed by the stream. The flat, undisturbed layer which once extended over the latter flow is visible in the background of the picture. The stream has modified the flow layer only by removing the least resistant, shaly portions. This view emphasizes the irregular structure of the flow.

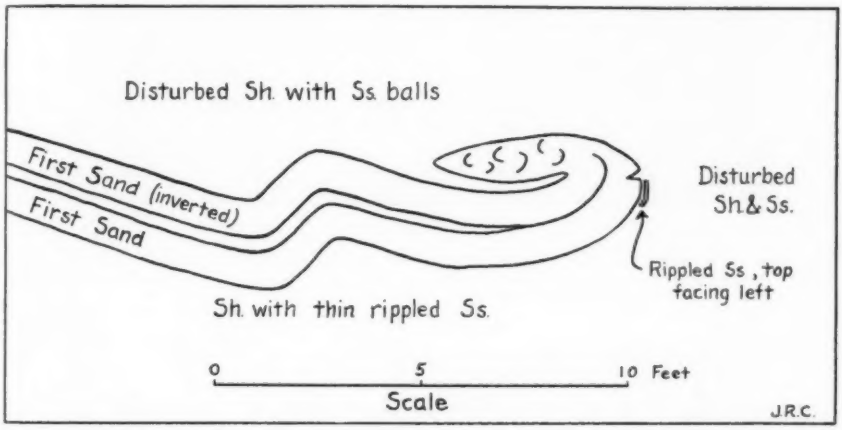

FIG. 4.—Field sketch of the relationships shown in Figure 3

There is a relationship between the structure of the flow and the kind of rock. Where both shaly and massive beds are involved, the shale is thickened under anticlines and elsewhere thinned or even pinched off. The presence of shale between sandstone beds has favored flowage. A shale seam lies at the base of some flows, and it is believed that at least a film of shale is present at the base of most, if not all, flows.

Flows are found in close association

FIG. 5.—Typical flow layer, 4 feet thick, in the Berea sandstone, long cliff at Rocky Fork. The total thickness of beds shown is approximately 30 feet. The flow layer appears near the middle of the cliff. (Photo by L. G. Westgate.)

FIG. 6.—Flow layer in the Berea sandstone exhumed by stream erosion, east-end section, Rocky Fork. Photo by L. G. Westgate.)

with gentle cross-bedding, as well as scour-and-fill channels. In spite of considerable search, no evidence was found that the slope of the sea bottom indicated by channels and inclined layers was the cause of any flow.

SUNBURY LOCALITY

Sandstone flows are well exposed in the Berea sandstone at Sunbury. The best place to study them is on the west side of Big Walnut Creek below the highway bridge just north of the village. The section here is shown in Figure 7. Several kinds of structure may be distinguished.

SHALE BOILS

In the bed of the creek and exposed only at very low water, the flat-lying rock layers are locally punched upward in boil-like forms—a kind of disturbance not recognized elsewhere. They were seen on the writer's first visit to the locality but have not been exposed for critical study since. It is therefore impossible to

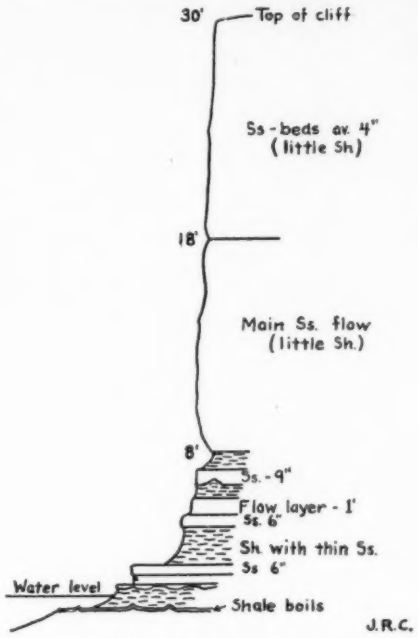

FIG. 7.—Section of Berea sandstone cliff at Sunbury.

FIG. 8.—Shale rolls in the base of a sandstone bed, Sunbury

give a detailed description of them at the present time.

SHALE ROLLS

At numerous places near the base of the cliff, beds of shale send irregular, rolling protuberances into the base of overlying sandstone beds (Fig. 8). The lamination of the shale is folded upward in the rolls, but bedding planes were not observed in the sandstone. In every case examined the top of the sandstone is flat, even though the shale rolls may nearly pierce it.

FLOW-FOLDS

At one place near the base of the cliff, flow-folds are clearly shown in a 10-inch zone of sandstone with shale partings (Fig. 9). One massive bed maintains essentially the same thickness throughout the folds, whereas the shaly layers are notably thickened under the anticlines. The folded zone is sharply truncated by flat-lying beds at the top but grades into undisturbed beds below.

MAIN SANDSTONE FLOW

This is a convenient name for the prominent disturbed zone shown in Figures 10, 11, and 12. It is 8–12 feet thick and 225 feet long. Winchell described it as "massive, or concretionary and irregular, with oblique divisional planes. . . ."[6] It is easy to see why the term "concretionary" was applied. The divisional planes do form rude circular and oval patterns repeated at intervals along the face of the cliff as though there were a series of large concretions. On critical study, however, it becomes apparent that none of the concentric structures is complete but that the lower half is invariably better developed than the upper. Thus the structures approximate more nearly semicircles having their open concave side facing upward. In conformance with this generalization the base of the whole layer is scalloped, whereas the top is essentially straight and horizontal. The scallops in the base appear to have been caused by the intrusion of the underlying shale between the pseudo-concretions.

It is not known what the structures are like in a direction at right angles to the cliff. They might and probably would appear quite different.

KNEADED SANDSTONE LAYERS

Some portions of the sandstone have a weathered surface resembling kneaded dough (Fig. 13). The structure is not always so apparent in the fresh rock, but even there crumpled bedding planes and breccia structure are in places discernible. The kneaded layers occur with unkneaded in several different kinds of flow structure. They also occur alone, as shown in Figure 13. In this instance the rock looks as though it had flowed with a rotary motion about numerous scattered centers. The top of the kneaded layer is truncated by a thin ripple-marked sandstone.

FLOWS CONTEMPORANEOUS

Two main problems were advanced for solution at the beginning of this paper, one concerned with dating the disturbances, the other with their cause. The first must be solved before the second is considered. The writer believes that the first problem is solved, that the disturbances took place on the surface of deposition before the overlying beds were laid down—that is, that there were contemporaneous mud and sand flows on the sea bottom. The evidence for this, repeatedly pointed out in describing the localities, is that the tops of the flows are truncated and overlain by undisturbed

[6] *Op. cit.*, p. 279.

FIG. 9.—Flow-folds in sandstone with shale partings at Sunbury

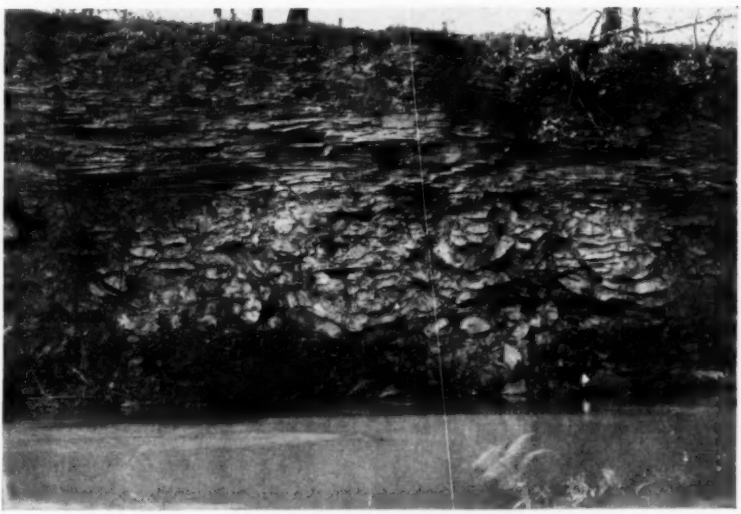

FIG. 10.—Main sandstone flow at Sunbury. A shale roll with intrusive tongues is visible at the base of the flow at one place.

beds. Shearing may be present at the bottom of the disturbed mass, but the top is a local unconformity and not a shear plane. Most flows were considerably eroded before deposition was renewed. This conclusion, based on the structural relationship of the flows themselves, is in accordance with other evidence bearing on the conditions of sedimentation. Erosion channels, cross-bedding, and ripple marks (mostly of the wave but some of the current type) prove turbulent water and alternate scour and fill.

CAUSE OF THE FLOWS

Although there is a considerable amount of literature on the contemporaneous flowage of subaqueous sediments, the details of movement and their causes are not so well known as those of mud flows and solifluction on the land. The controlling physical laws are the same in the two cases, but it must be remembered that the subaqueous sediment is not subject to occasional soaking with water and never rests on a frozen substratum. Moreover, the surface slope of the sediment is generally less than that

FIG. 11.—Shale squeezed in between pseudo-concretions of sandstone at the base of the main sandstone flow at Sunbury.

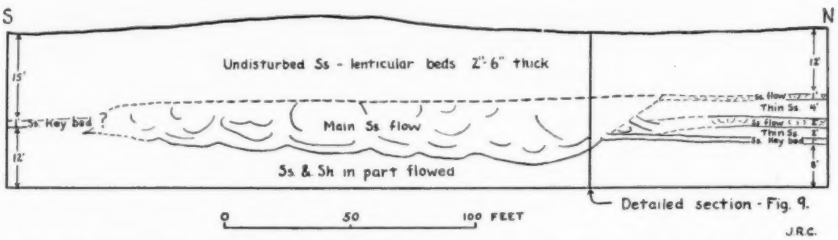

FIG. 12.—Structural relationships of main sandstone flow at Sunbury. Vertical scale is twice horizontal

of the land. On the other hand, freshly deposited sediment is highly charged with water and thus internal friction is less. Therefore, less force is necessary to muds, especially under water, that slipping may be induced even on slopes of only 2° or 3°."[7] It has been known to occur on a slope of 2°31′.[8]

FIG. 13.—Kneaded sandstone truncated by a thin, ripple-marked layer, old quarry at Sunbury

dislodge it. The following factors are involved:

1. Surface slope
   a) Deposition on sloping surface
   b) Irregular deposition
   c) Irregular compaction or removal of supporting matter below
   d) Contemporaneous scour
   e) Diastrophic movements
2. Composition of sediments
3. Decreased friction in the mass
4. Weight of sediments
5. Outside impulses

SURFACE SLOPE

A slight inclination in the surface of newly deposited, soft sediments may be sufficient to cause a flow or a slide. All that is necessary is that the component of gravity in the direction of slope exceed the internal friction in the sediment. According to F. H. Lahee: "The coefficient of friction is so low in moist clays and

Surface slope is considered the fundamental cause of the Bedford and Berea flows, although it cannot be proved that any particular flow was caused thereby. This is not surprising, since initial slopes "of only 2° or 3°" would not be determinable in an irregularly bedded formation subsequently tilted out of the position of deposition.

a) *Deposition on sloping surface.*—The surface of recent sediments in seas and lakes is generally undulating, with slopes up to a number of degrees.[9] The offshore slope is a familiar example. Deposition on broad regional slopes of this kind may have been a factor in causing the Berea

[7] *Field Geology* (New York: McGraw-Hill Book Co., 1941), p. 192.

[8] W. H. Twenhofel, *Treatise on Sedimentation* (Baltimore: Williams & Wilkins, 1932), pp. 739–40.

[9] *Ibid.*, pp. 604–5.

and Bedford flows. It is impossible to evaluate this factor, but local irregularities in the surface of deposition due to nonuniform thickness of older beds and contemporaneous erosion are evident. Layers of proper consistency deposited on such a surface might flow and thus form the observed structures.

b) Irregular deposition.—The foreset beds of a delta, sand bar, or other form of outbuilding sediment may flow under proper circumstances. Gentle cross-bedding is characteristic of the Berea, and the slopes indicated must be considered a possible cause of flow.

c) Irregular compaction or removal of supporting matter below.—Level deposits may be tilted by unequal compaction, by solution of a buried soluble layer, or by the squeezing-out laterally of mobile layers below. The mud lumps of the Mississippi delta are a familiar result of the last process.[10] There is no soluble layer for at least 600 feet beneath the Berea and Bedford formations of central Ohio, but the other processes may have operated. The Bedford is a soft clay shale. When freshly deposited it must have been highly compactible and mobile under stress. Outcrops near the base of the Berea along Big Walnut Creek at Sunbury show bodies of massive sandstone in the Bedford shale as if the Berea had sunk into it. Stauffer gives the thickness of the Berea as 64 feet, 8 inches at this locality,[11] to be compared with a usual thickness of about 40 feet. About it he writes:

> . . . . The Bedford shale decreases in thickness to the northward. By the time the Sunbury region is reached an even greater decrease is noticed, while the simultaneous thickening of the Berea grit is a feature of the northward sections.[12]

This is the relationship that would be expected if the Bedford had been squeezed out from under an extra-heavy load in the Sunbury area. This is pointed out, since it may explain the structures of those particular outcrops, not because it is considered a significant cause of other flows.

d) Contemporaneous scour.—Sufficient slope to cause a flow or slump may result from the contemporaneous erosion of channels in level sediments. The large slump exposed in the tree cliff on Rocky Fork is best explained in this way. Many examples of contemporaneous erosion channels are found in the Berea proper, but these are all shallow and have no apparent causal relationship to flows.

e) Diastrophic movements.—Level sediments may be tilted by regional warping of the earth's crust, but there are no indications that such movements were involved in the Berea and Bedford flows.

COMPOSITION OF SEDIMENTS

Field observations and laboratory experiments show that muds are more subject to flow than sands. In any theory that seeks to explain the Berea sandstone flows, the composition of the rock must be considered. The Berea is not a clean, well-washed sand. There is considerable argillaceous material mixed with the sand and also as separate beds and seams. Unquestionably this acted as a lubricant. A clay seam is commonly found at the bottom of the flowed portion.

DECREASED FRICTION IN THE MASS

This is one of the most important causes of subaerial solifluction but sel-

[10] E. W. Shaw, "The Mud Lumps of the Mouths of the Mississippi," *U.S. Geol. Surv. Prof. Paper 85-B* (1913).

[11] "Geology of the Columbus Quadrangle," *Geol. Surv. Ohio*, Bull. XIV (4th ser., 1911), p. 33.

[12] *Ibid.*, p. 31.

dom the condition of subaqueous flows. As Assar Hadding points out:

> The changes occurring in deposited sediments (the diagenesis) are as a rule a hindrance to movements of the mass. The compression, dehydration, crystallization and cementation—all act in the same direction. In subaquatic sediments there is no correspondence to the subaerial sediments' occasional saturation with water, frost in the ground, and thaws with subsequent solifluction phenomena.[13]

WEIGHT OF SEDIMENTS

Hadding clearly points out the importance of this factor in causing subaqueous slides.[14] If the surface of deposition is slightly inclined, friction may be sufficient to hold a thin and therefore light layer of sediment in place but insufficient to hold a thicker, heavier mass. A spontaneous flow or slide will result after the accumulation of a certain optimum thickness, depending on the characteristics of the sediment and the degree of slope. Unctuous clay seams in the mass will favor movements by providing planes of easy shear. If a formation like the Berea has flows within it, we should expect to find them bottomed by clay seams.

OUTSIDE IMPULSES

Subaqueous flows and similar-looking disturbances have been caused by earthquakes, the grounding of icebergs, and other outside impulses. There is no reason to suppose that the structures in the Bedford shale and Berea sandstone were caused in this way.

ACKNOWLEDGMENT.—I wish to acknowledge the help received from Professor Lewis G. Westgate, of Ohio Wesleyan University, in the preparation of this paper. He encouraged me to undertake the study in the first place, gave freely of his time to advise and discuss as the work progressed, and has permitted the use of his photographs. Of course, I assume complete responsibility for the interpretations and conclusions.

[13] "On Subaqueous Slides," *Geol. fören. i. Stockholm förhandl.*, Vol. LIII, Part IV (1931), p. 379.

[14] *Ibid.*, pp. 378–79.

SOIL AND SUBSEQUENT TOPOGRAPHY[1]

P. E. WOLFE
Princeton University

ABSTRACT

Fluvial erosion acting on a series of plunging synclines and anticlines has etched out a subsequent topography the details of which correspond intimately with structural, lithologic, and soil variations. A fine-textured erosional pattern on the dip slopes as contrasted with a coarse-textured pattern on the scarp slopes of the strike valleys is explained on the basis of soil-erosibility differences, which in turn are due to lithologic and structural differences. Structural lines of weakness such as faults and fractures in the crests of anticlines have localized valley development. Tributaries to these anticlinal valleys have developed along minor faults, fractures, and soft shale beds. The synclinal interfluves which are composed of alternate beds of hard and soft shale have been eroded to a strike-vale-and-ridge topography the details of which are attributed to structural, lithologic, and soil-erosibility differences. This strike-vale-and-ridge topography is post-Kansan and pre-Wisconsin in age.

INTRODUCTION

Detailed geologic and soil studies were carried out in two areas of the Piedmont Triassic lowlands, one at Frenchtown, New Jersey, and the other at East Greenville, Pennsylvania. These studies showed that the land forms in both areas were excellent examples of a subsequent topography formed by erosional processes and that structure, lithology, and soil all played important roles in the development of the land forms.

LITHOLOGY AND STRUCTURE

The bedrock in both of the areas studied is the Brunswick formation of Triassic age. This is composed of alternating beds of soft and hard red shales and of a gray-blue argillitic shale. In the East Greenville and northern part of the Frenchtown area these shales are thinly bedded with alternating beds ranging in thickness from 2 to 10 feet. In the southern part of the Frenchtown area the beds are more massive and range in thickness from 10 to 100 feet or more. These shales were folded in late Triassic time into a large westward-plunging syncline within which there are smaller synclines and anticlines. The average dip of the beds ranges from 2° to 10°. Joint systems are as a rule well developed, and the bedrock is broken in places by minor faults of small displacement.

DESCRIPTION OF TOPOGRAPHY

The present topography is a maturely dissected plain or lowland which is made up of the following elements:

1. Strike valleys and uplands which have been developed on the south limb of the major westward-plunging syncline. The sides of these strike valleys are further characterized by a type of topography in which (*a*) the dip slopes have a fine-textured erosional pattern with 1–2-foot gullies spaced 25–50 feet apart, whereas (*b*) the scarp slopes have a coarse-textured erosional pattern with 25–50-foot deep, round-headed depressions spaced 500–800 feet apart. These round-headed depressions are termed "dales"[2] in this paper. The accompanying diagrammatic sketch (Fig. 2) illustrates these features.
2. Narrow anticlinal valleys and broad synclinal interfluves developed on the smaller folds within the major syncline. The synclinal interfluves have a distinctive topography which consists of 4–10-foot strike

[1] This article has been condensed from a dissertation submitted to the faculty of Princeton University as a partial requirement for the degree of Doctor of Philosophy.

[2] Usage of term "dale" suggested by C. W. Thornthwaite in a personal communication for broad round-headed depressions, the sides and bottom of which merge without any abrupt angle.

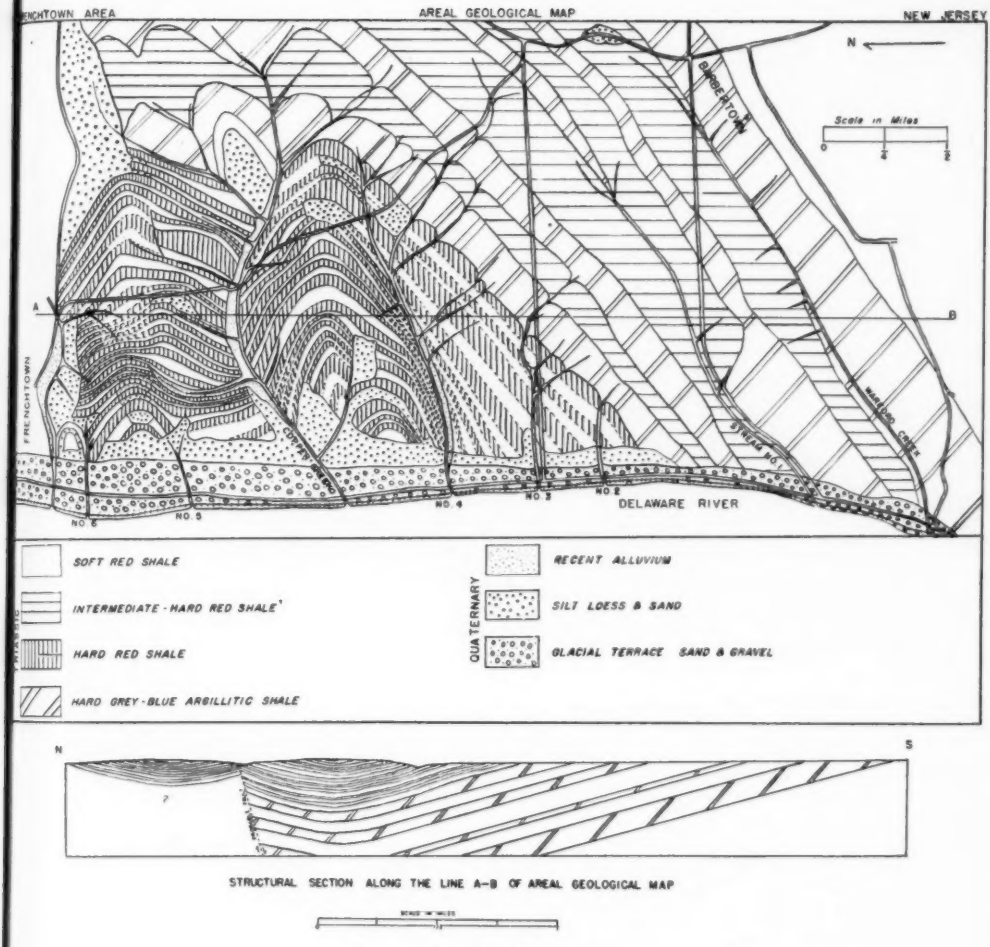

FIG. 1.—Areal geological map

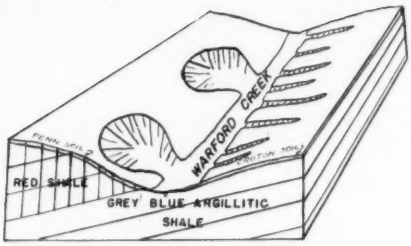

FIG. 2.—Diagrammatic sketch showing erosional pattern differences on scarp and dip slopes of strike valley.

ridges and valleys spaced 50–200 feet apart. Because of the small dimensions of these ridges and valleys, this type of land form has been termed a "strike-vale-and-ridge" topography in this paper, the term "vale" having been previously used by S. W. Woolbridge and R. S. Morgan[3] to indicate very shallow strike valleys. The accompanying contour map (Fig. 3) and Figure 4 illustrate these topographic features.

DEVELOPMENT OF TOPOGRAPHY INFLUENCED BY STRUCTURE, LITHOLOGY AND SOIL

INFLUENCE OF STRUCTURE

The narrow anticlinal valleys and broad synclinal interfluves which occur in the East Greenville and northern part

[3] *The Physical basis of Geography* (*Geomorphology*) (London and New York: Longmans, Green & Co., 1937), p. 203.

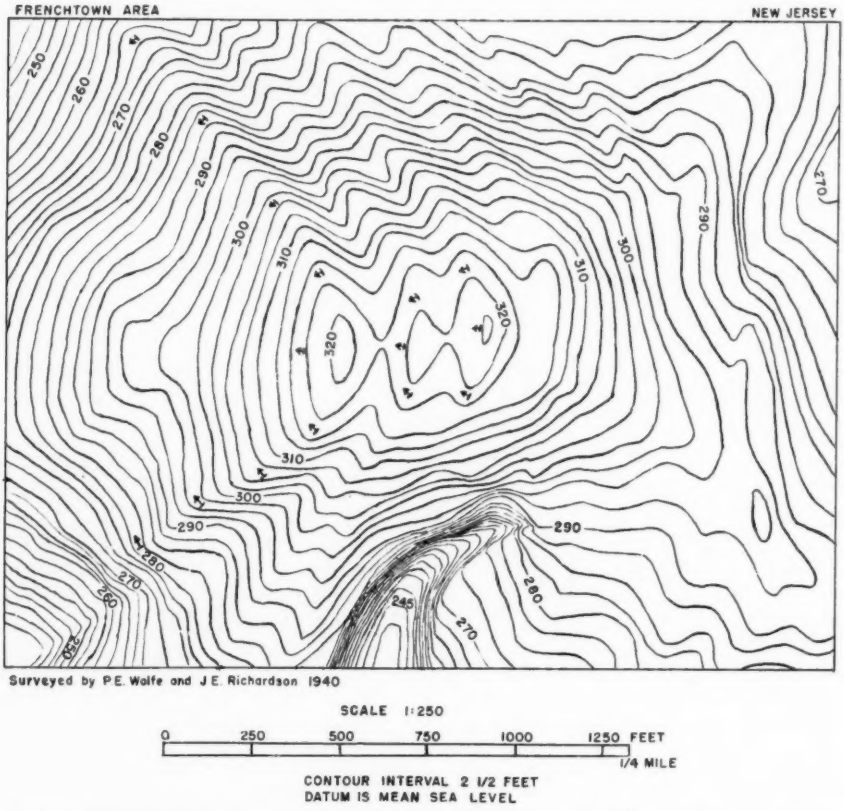

FIG. 3.—Contour map showing strike-vale-and-ridge topography of the area in the northwest corner of the map, Fig. 1.

of the Frenchtown areas have been markedly controlled by the structure of the plunging folds. These folds are of the parallel type, and during peneplanation they were beveled to such a level as to leave broad synclines and narrow anticlines. Following peneplanation the area was uplifted, and a new cycle of erosion was begun in which the truncated anticlines were eroded more rapidly than the synclines so that a synclinal-upland and anticlinal-valley topography was produced. The presence of faults and of fractures in the crests of anticlines was important in localizing valleys. Tributaries to these anticlinal valleys also develop along structural lines of weakness such as faults and zones of tight folding where the rocks were badly fractured.

The strike-vale-and-ridge topography developed on the synclinal interfluves is also structurally controlled. Here differential erosion of the soils mantling the hard and soft shale beds, as will be brought out later, has etched out a subsequent topography the details of which correspond intimately with the strike of the beds of the plunging synclines. In many places this strike-vale-and-ridge topography has been further dissected by steep-walled tributaries which are centered in zones where the strike of the beds changes abruptly and the shale beds have been badly fractured.

INFLUENCE OF LITHOLOGY AND SOIL

Because lithology indirectly through the medium of the soil has been found to be important in erosional land-form development and also since the soils in both the Frenchtown and East Greenville areas are endodynamomorphic or skeletal soils and therefore closely related to the underlying geologic materials, these two factors are discussed together. As stated earlier, in the southern part of the Frenchtown area the dip slope of the strike valleys has a fine-textured, corrugated, erosional pattern with many closely spaced shallow gullies. This slope is underlain by a gray-blue argillitic shale mantled by the Croton soil.[4] This soil has a dense clayey hardpan *B* horizon which is impermeable to water, and, consequently, the infiltration rate is low and the immediate runoff high. The hardpan horizon is quite resistant to erosion, so that, once a gully cuts through the *A* horizon, farther down cutting is greatly retarded. Because of this the gullies are shallow, and the hardpan horizon rather than bedrock is exposed in the gullies.

FIG. 4.—Strike-vale-and-ridge topography showing coarse shale rubble on strike ridge.

On the scarp slope of these strike valleys the coarse-textured erosional pattern with few widely spaced dales is also attributed to the influence of lithology and soil. This slope is underlain by the soft red shale which is mantled by the porous well-drained Penn soil.[5] This soil has a higher infiltration and lower immediate runoff rate than the dip-slope soils, and, since it lacks the dense clayey hard-

[4] More complete descriptions of these soil types may be found in A. L. Patrick *et al.*, "Soil Survey of the Belvidere Area of New Jersey," *U.S. Dept. Agric. Conser. Development New Jersey* (1902), pp. 49–53, 58–60.

[5] *Ibid.*

pan horizon characteristic of the Croton soil, gully erosion is not retarded when cutting has progressed through the *A* horizon but continues through the *B* horizon to the underlying weathered bedrock, thus resulting in broad deep gullies.

In the East Greenville area a series of pits dug over one of the strike-vales-and-ridges showed that there are sharply differentiated soils corresponding to the two different underlying lithologic types. The contact of the soil mantle and underlying shales is gradational, the soil mantle grading up from fresh parent-rock into weathered parent-material or *C* horizon, and this into the overlying solum. The Penn shale loam mantles the soft red shale of the strike-vales, while the Penn gravelly loam overlies the hard shale ridges. The first of these soils consists of 7–10 inches of an Indian-red silt loam *A* horizon which overlies 8–10 inches of a brighter Indian-red clay loam of the *B* horizon. This, in turn, overlies weathered parent-shale which grades downward into fresh soft red shale. The Penn gravelly loam, or, as it might be more appropriately called, "shale rubble," mantles the strike ridges. This mantle is made up of 8–12 inches of mixed *A*, *B*, and *C* horizons which consist of an Indian-red silt loam mixed with large quantities of coarse shale fragments. Tillage of the thin-soil mantle on the hard shale ridges has undoubtedly been responsible for the mixing of the *A*, *B*, and *C* horizons. This material in turn grades down into weathered shale which overlies fresh hard shale bedrock. Microscopic studies of these shales showed that the hard red shale had a much higher siliceous cement content than the soft red shale. This can be seen in the photomicrographs (Figs. 5 and 6). It is thought that because of the higher siliceous content the hard red shale weathers to larger fragments than the soft shale and thus gives rise to a shale rubble or gravelly soil. In each case it is the soil mantle and not bare rock that is being eroded, and the present strike-vale-and-ridge topography is a result of the differential erosibility of the soils which mantle the hard and soft red shales; the soil type, however, is directly related to the underlying lithology. Thus lithology indirectly through the medium of the soil has been important in the development of this topography.

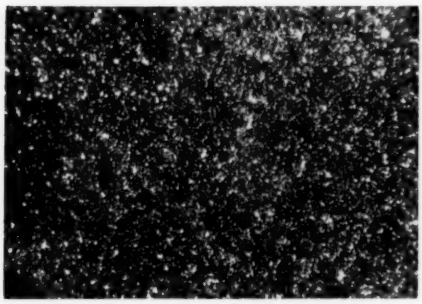

FIG. 5.—Photomicrograph of soft red shale (3 × crossed nicols).

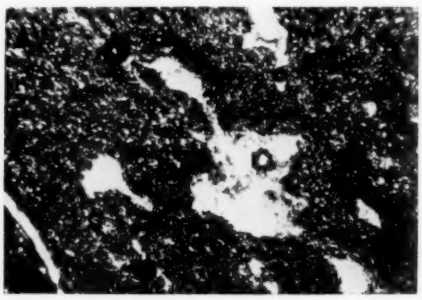

FIG. 6.—Photomicrograph of hard red shale showing silica as white masses (3 × crossed nicols).

In the northern part of the Frenchtown area a trench dug to bedrock over one of the strike-vales-and-ridges along

the axis of the syncline at the watershed divide revealed the presence of loess lying directly on bedrock. This loess was up to a foot or more in thickness in the vales and thinned out over the ridges. The contact of the loess and underlying rock was sharp, thus being in marked contrast to the soils of the East Greenville area in which the contact graded from the solum to weathered parent-material to fresh bedrock. Furthermore, microscopic studies made of loess and underlying shales showed a distinct difference in grain size, bringing out the fact that the loess material could not have been derived from the weathering of the shale. Inasmuch as the loess is coarser grained and sandy nearer the Delaware River and finer grained inland, it seems clear that the immediate source of the material was the Wisconsin terraces bordering the river.

The fact that the loess deposits lie on fresh bedrock rather than on a fossil soil indicates that pre-loess erosion was severe and that the soil mantle was stripped prior to the deposition of the loess. The unusual conditions of erosion may well have been determined by the tundra climate of the glacial period, sparse vegetation, and severe wind and fluvial erosion.

AGE OF STRIKE-VALE-AND-RIDGE TOPOGRAPHY

The discovery of eolian silt and sand deposits lying on fresh bedrock in the strike-vales and the fact that the strike-vales-and-ridges pass under thick deposits of wind-blown material served as a convenient method for dating the topography in the Frenchtown area. Since the wind-blown silt and sand is of Wisconsin age, the upper age limit of the strike-vale-and-ridge topography is pre-Wisconsin.

The lower age limit of the strike-vale-and-ridge topography is based on the age of the peneplain which beveled the folded rocks of the area and in which the strike-vales-and-ridges were subsequently

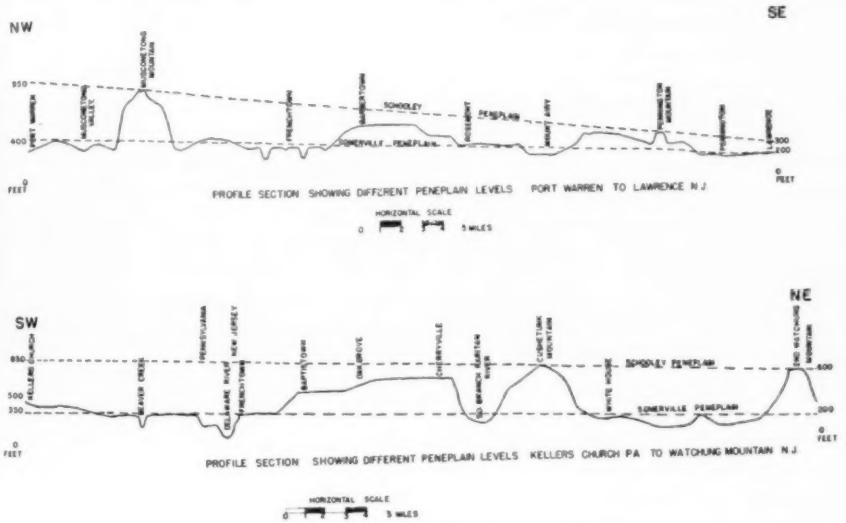

FIG. 7.—Profile sections showing different peneplain levels

etched out. In determining the peneplain surface represented in the Frenchtown area, profile sections were made from topographic maps and known peneplain levels were connected so that the intervening areas might be correlated. Two profile sections were made. One of these extends from Port Warren to Lawrence, New Jersey, and is approximately at right angles to the New Jersey pre-Cambrian highlands. The levels at Lawrence and Pennington at one end of the section have been classified as the Somerville peneplain by W. M. Davis[6] and others. A line connecting these known points indicates that the accordant summit levels of the Frenchtown area fall within the bounds of the Somerville peneplain level (see profile sections, Fig. 7). The profile section parallel to the New Jersey highlands from Kellers Church, Pennsylvania, to Second Watchung Mountains, New Jersey, similarly shows that the Frenchtown area falls within the Somerville peneplain level. From the evidence of these profile sections plus field observations which show that the Frenchtown surface is considerably lower than the New Jersey highlands (Schooley peneplain) it has been deduced that the accordant summit levels of the Frenchtown area are remnants of the Somerville peneplain.

The determination of the age of this peneplain surface is based on the work done by R. D. Salisbury,[7] who concluded that Jerseyan drift was deposited on the Somerville peneplain prior to its uplift and dissection. Later Frank Leverett[8] and P. MacClintock[9] divided the Jerseyan glacial drift into two drift sheets. The older has been correlated with the Kansan and the intermediate drift sheet with the Illinoian. Since the older of these drift sheets (Kansan) was deposited on the undissected Somerville peneplain surface, the age of the peneplain is assumed to be Kansan. Hence the strike-vale-and-ridge topography is post-Somerville peneplain (Kansan) and, from what has been said earlier, pre-loess (Wisconsin) in age.

[6] "The Geographic Development of Northern New Jersey," *Proc. Boston Soc. Nat. Hist.*, Vol. XXIV (1890), p. 365.

[7] *N.J. Geol. Surv. Ann. Rept.* (1893), p. 76.

[8] "Glacial Deposits outside of the Wisconsin Terminal Moraine in Pennsylvania," *Pa. Geol. Surv. Bull. G7* (4th ser., 1934), pp. 78–90.

[9] "Weathering of the Jerseyan Till," *Bull. Geol. Soc. Amer.*, Vol. LI (1940), p. 113.

SUMMARY AND CONCLUSIONS

1. Fluvial erosion acting on a series of folded rocks in the Frenchtown, New Jersey, and the East Greenville, Pennsylvania, areas of the Piedmont Triassic lowlands has etched out a subsequent topography, the details of which correspond intimately with structural, lithologic, and soil variations.

2. Differential erosion of lithologic units in a humid climate is in reality accomplished by the erosion of the soils which mantle the respective units rather than of bedrock itself. It is only under limited circumstances, such as streambed channels and some gully beds, that bedrock itself is eroded.

3. The fine-textured corrugated erosional pattern on the dip slope as contrasted with the coarse-textured widely spaced dale pattern on the scarp slope of the strike valleys is accounted for by soil-erosibility differences, which in turn are due to lithologic and structural differences.

4. The broad synclinal interfluves and narrow anticlinal valleys have formed by peneplanation of plunging parallel folds and by the subsequent uplift and sculpturing during the present erosion cycle.

Structural lines of weakness such as faults and the concentration of fractures in the crests of anticlines have been important in the development of anticlinal valleys.

5. Tributaries to the main anticlinal valleys develop along the strike of soft shale beds, along faults, and along badly fractured areas where tight folding and bending have shattered the shale beds.

6. The synclinal interfluves have been etched out to a strike-vale-and-ridge topography the details of which correspond intimately with structural, lithologic, and soil variations.

7. The strike-vale-and-ridge topography is post-Somerville peneplain (Kansan) and pre-loess (Wisconsin) in age.

ACKNOWLEDGMENTS.—The author is indebted to the department of geology, Princeton University, and to the Climatic and Physiographic Section of the Soil Conservation Service for the funds which made this study possible. Sincere thanks are also extended to the members of the department of geology and especially to Dr. Paul MacClintock for advice, suggestions, and criticisms during the course of field study and preparation of the manuscript and also to Drs. C. W. Thornthwaite and L. L. Lee of the Soil Conservation Service for advice during several field conferences.

FRANK DAWSON ADAMS
1859–1942

Canada's foremost geologist has rounded out the full span of a long, unusually fruitful life devoted unselfishly to science and public service. His death on December 26, 1942, in his eighty-fourth year, brought to its close a career rich alike in scientific achievements, human betterment, and official duties admirably performed. The gap which follows will be hard to fill.

Dr. Adams' published contributions have treated many aspects of geology, reflecting the broad interests of the man. But quite naturally the strong geologic appeal of his regional environment, during three-score years at McGill University in Montreal, directed his chief attention to the pre-Cambrian rocks of the Canadian shield and perforce led him into the fields of petrography, structural geology, and economic geology. The neighboring Laurentian area became the scene of much of his field work, and, as his studies progressed, the inherited interpretations of its very ancient history underwent profound modifications. The last half-century has witnessed no more radical revision of any portion of interpreted North American geological history than the truer and fuller deciphering of its obscure early chapters in which Adams took a leading part.

For many years Dr. Adams carried on that succession of fundamental laboratory researches on the deformation of rocks at high pressures and temperatures for which his name will long be known. *The Flow of Marble* in 1910 was a milestone of progress in earth dynamics. Ensuing publications on the behavior of other rocks under strong confining pressures and high differential stress at varying temperatures led to new concepts of the so-called "zone of flow" and aided materially in understanding the nature of rock folding and the intimate relationship between fracture and flow. The chastening effects of quantitative experimental results were now brought to bear on previous deductions from less certain premises.

He traveled widely. After his retirement appeared *The Geology of Ceylon*, which may be cited to indicate the far reach of his constructive efforts. For his last major contribution, *The Birth and Development of the Geological Sciences*, he visited may of the leading libraries of Europe and himself collected a remarkable library on the history of geology and closely related sciences which he presented to McGill University shortly before his death.

He was in all respects a very modest man, having the simple dignity and quiet courtesy of the true gentleman with a great capacity for strong and enduring friendship. Both time and hearty support were given unsparingly to numerous altruistic activities wherein his humanity and sympathy found free vent. Honors came frequently, but only four will be mentioned here. He was a fellow of the Royal Society of London, was the only Canadian foreign associate of our National Academy of Sciences, served as president of the twelfth International Geological Congress in 1913, and was president of the Geological Society of America in 1917.

ROLLIN T. CHAMBERLIN

REVIEWS

Economic Mineral Deposits. By Alan M. Bateman. New York: John Wiley & Sons, Inc., 1942. Pp. vii+898; figs. 300. $6.50.

The author of this latest American textbook on applied geology speaks from the rich background of an unusually varied experience as professor of economic geology at Yale University and as a successful consultant, an experience which he is now applying as chief of the Metals and Minerals Division of the Board of Economic Warfare.

While the book is of moderate size, its range includes almost the entire field of applied geology—a range that few authors have had the temerity to enclose within the covers of a single volume. The Preface states: "Its chief purpose is as an elementary textbook but it could be adapted almost equally well to more advanced courses. It is meant to be a source of information also to all those interested in the mineral industries." The book is well proportioned and the condensation discriminatingly done; the illustrations are numerous and well chosen, and most of them are new. However, for advanced students and professional geologists it is of particular interest and importance to know the sources that have been consulted in the preparation, especially of regional and district summaries. The omission, obviously for brevity, of all footnote references to original sources will force such readers to other source books for such references. The brief list of selected references at the end of sections does not compensate for the lack of full footnotes.

Part I, which constitutes nearly half of the volume, is devoted to principles and processes; Part II, comprising about a fourth of the volume, relates to metallic mineral deposits; and Part III, comprising the final fourth, to nonmetallic mineral deposits.

Part I develops the principles underlying the formation of both metallic and nonmetallic resources, including the mineral fuels. A noteworthy feature of this part of the book is the section on the stability of minerals, including a two-page list of inversion points, melting points, and exsolution temperatures, that may contribute definite information with regard to the temperatures of mineral formation. In the 237 pages devoted to the processes of formation of mineral deposits the classification is partly genetic, making use of such categories as "magmatic concentration" and "contact metamorphism," and partly descriptive, with such categories as "metasomatic replacement" and "cavity filling." This mixed basis of classification is proposed primarily as a working classification for the use of beginning students in economic geology and for the mining geologists and mine operators. In using such a classification, justifiable as it may be on the grounds of simplicity, care must be taken even with beginners to make clear such fundamental facts as the metasomatic nature of contact metamorphism and the fact that metasomatic replacement may take place throughout most of the wide range of mineral formation from pegmatite to supergene enrichment.

The magmatic deposits are classified into early and late magmatic with subdivisions of each; and this is a forward step. Pegmatites are classified under the late magmatic deposits as "residual liquid injections," where they are somewhat incongruously teamed up with the magmatic deposits of Kirruna and the Adirondacks. As it amply deserves from its importance as a mechanism of ore deposition, replacement is treated well and at length. Most of the structural features of mineral veins, including ore shoots, are described at considerable length in the section on "Cavity Filling." Others are considered in chapter vi under "Controls of Mineral Localization" and in chapter vii on "Folding and Faulting of Mineral Deposits."

In appreciation of the work of Locke, Blanchard, and Boswell the mineral and textural features of gossans as indicators of the ores at depth are well presented in the section on oxidation and supergene enrichment.

Inclusion in a text primarily designed for geologists of a chapter on "Geology and Prospecting Exploration Development and Valuation of Mineral Properties" and another chapter on "Extraction of Metals and Minerals" is an altogether desirable addition, since in many schools geologists receive little, if any, course work in these primarily engineering subjects.

In Part II metals are considered seriatim,

the treatment being mainly summary descriptions of leading occurrences grouped geographically, supplemented by very minor treatment of history, production tenor, and uses.

Part III is devoted to nonmetallic deposits, including coal, and petroleum and natural gas and ground-water supplies. It contains a wealth of recent material not found in other treatises on nonmetallics and thus reflects the great progress in recent years in the industrial-minerals field.

E. S. B.

Pennsylvanian System in New Mexico. By M. L. THOMPSON. (New Mexico School of Mines Bull. 17.) Socorro: State Bureau of Mines and Mineral Resources, 1942. Pp. 92; figs. 8; pls. 2. Price $0.50.

This much-needed report is a precursor to a more extensive work to be published later. Its purpose is to "set up a basis for a detailed classification of the Pennsylvanian of New Mexico." In this the author has been admirably successful.

The report begins with a concise general treatment of the geographic distribution, lithology, thickness, and paleogeography of the Pennsylvanian system, followed by discussions of the Pennsylvanian and post-Pennsylvanian land masses and basal Pennsylvanian and Permian contacts. There is also a brief but apparently complete synopsis of previous stratigraphic classifications accompanied by a chronologic chart. The paleogeographic section based on all available information impresses the reader as one of the outstanding contributions of this report.

Owing to the difficulty in correlating the limestone beds, most students have hitherto considered the Pennsylvanian of New Mexico as comprising not more than two formations. But now, through extensive field work, faunal investigation, and conferences with other Pennsylvanian workers, the author has brought out a much more detailed division of the Pennsylvanian system and proposes a new set of series, group, and formational names. Four series are recognized—the Derry, Des Moines, Missouri, and Virgil. The new series name—Derry—is applied to the oldest Pennsylvanian in New Mexico, and the author considers its introduction essential because the beds represented are younger than Morrow and older than Des Moines.

Under each of the four series the author has erected two new groups, and there are sixteen new formations and members. Each is treated separately, and type sections are set forth with complete descriptions of the geographic occurrence, lithology, thickness, individual fossils, and fossil zones.

Fusulinid zones are the chief basis for the various divisions and likewise for correlations with areas outside New Mexico. The author decides that, because "the lithology of the Pennsylvanian of New Mexico varies so markedly geographically, especially in its lower and upper portions, not all lithologic units in New Mexico will be named. . . ." But, despite the fact that this report necessarily is not complete in all detail, it will afford the reader an excellent outline of the Pennsylvanian system of New Mexico.

FRANK V. STEVENSON

Oil Property Valuation. By PAUL PAINE. New York: John Wiley & Sons, Inc., 1942. Pp. viii+204; figs. 34. $2.75.

An area of active "wildcat" drilling is chiefly of interest to the geologist. When oil is found, the area which constitutes the new field presents more than geological problems. Production becomes established, reserves are computed, and the property (as proved and semiproved land) takes on a new value. This is treated in *Oil Property Valuation.* It is not geology, but the work deserves reading by geologists because it deals with a collateral problem.

The book is comprehensive, clear, and concise. The beginning chapter, "Scope of Valuation," tells of the meaning and general methods of valuations. The second chapter is on "Properties" and has descriptions of all types of oil properties and property interests. A separate chapter is devoted to unproved lands. "Oil and Gas Reserves" is particularly interesting to one not trained in petroleum engineering, for it sets forth the chief methods of arriving at estimates of reserves. The last three chapters—"Elements in Valuation," "Valuation Methods," and "The Examination and Report"—have to do with accounting and actual report-preparation techniques.

An Index and short bibliographies with each chapter add to the usefulness of the book.

C. W. STERNBERG

Among the articles to appear in early numbers of *The Journal of Geology* are the following:

Radiogenic Heat in Rocks. By NORMAN B. KEEVIL

Chink-Faceted Pebbles: Fluviatile versus Marine. By CHESTER K. WENTWORTH

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Friction Cracks and the Direction of Glacial Movement. By STANLEY E. HARRIS, JR.

Glacial Potholes on Outpost Islands, Great Slave Lake, N.W.T. By CHRISTOPHER RILEY

In the path of war and on a main supply route of the United Nations, the Fiji Islands assume new importance, and this book gives valuable information from firsthand survey.

FIJI: Little India of the Pacific

By JOHN WESLEY COULTER

Written by a geographer who has lived thirteen years in the south sea islands, this small book compresses much factual data on the geography, economics, and anthropology of the Southern Pacific area with an understanding of the people and their social psychology which affect the broad problems of international relations.

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